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Morphology and evolution of the Bay of Fundy, with emphasis on changes of sea level during the quaternary

MORPHOLOGY AND EVOLUTION OF THE BAY OF FUNDY WITH EMPHASIS ON CHANGES OF SEA-LEVEL DURING THE QUATERNARY

VOLUME I: TEXT

by

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Brandon, Manitoba March, 1971

MEMORANDUM

The work contained in this dissertation has not been submitted to Bristol University or to any other degree awarding institution for consideration for the awarding of a degree or diploma.

With the following exceptions the work is my own:

(1) Shell samples were identified by Dr. Frances J. E. Wagner, Atlantic Oceanographic Laboratory, Dartmouth, Nova Scotia.

(2) The programme for computing depth to bedrock using an Olivetti 101 calculator was devised by Dr. F. Hewitt, National Computing Centre, Manchester.

(3) Pollen analysis was carried out by Dr. K. Crabtree, Department of Geography, Bristol University.

Where reference is made to research carried out by other workers it is acknowledged in the text.

J. Welsted, March 23, 1971.

J. Welsted.

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Officers of the Maritime Marsh Rehabilitation Administration (M.M.R.A.) in Truro, Nova Scotia made available unpublished large scale maps of some of the marsh areas, and also the records from borings made in connection with the construction of dams across some of the rivers which drain the marshes. Similarly Warnock Hersey International Limited provided the writer with records of borings made in connection with the Amherst Bypass.

While doing field work the writer received able assistance from Mr. Michael Doiron (1965) who was at the time a Grade 12 student at Oromocto High School, New Brunswick; Mr. Leo Dare (1966) and Mr. Tom Carter (1967) who were at the time undergraduate students at Brandon University. Many undergraduates at Brandon University, too numerous to name individually, helped the writer with analysis of data collected in the field, with map analysis, and with air-photo interpretation.

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ABSTRACT

The Bay of Fundy is an arm of the Atlantic Ocean that almost separates Nova Scotia from New Brunswick, the two provinces being joined only by the low Chignecto Isthmus. The geometrical proportions of the Bay are such that it has the greatest known tidal range in the world.

Rocks exposed along the Fundy coast range in age from Precambrian to Recent, but there is a gap between the end of the Triassic and the Pleistocene. Variations in rock type, structure, exposure, and geomorphological history (including changes of sea-level) have combined to produce a great variety of coastal types, one of the most distinctive features of the Bay being the extensive tidal marshes and mud-flats.

A trough existed in the Fundy area by Triassic times but it was not below sea-level. Continental sediments were deposited in the trough with interruptions by volcanic activity. Later,major drainage channels were established along what are now the Bay of Fundy and the Annapolis - Cornwallis Lowland. Early workers in the area claimed that they could recognize in New Brunswick and Nova Scotia the remnants of a peneplain that was formed by the end of the Cretaceous. Although the idea persisted for many years and is still referred to occasionally, it is shown that there is no basis for believing that such a landform developed.

During the Pleistocene, New Brunswick and Nova Scotia were covered by an ice sheet which at its maximum reached the Scotian Shelf. During deglaciation the ice sheet split along the Bay of Fundy and its offshoots.

Glaciation and deglaciation resulted in changes of relative sea-level in the Bay. At times sea-level was higher than at present and at other times lower. Determination of exact changes of sea-level is complicated by the fact that tidal range in the past was not the same as it is now.

The evidence for emergence consists of shell-bearing clays and marine-formed morphological features now above sea-level. Emergence of at least 130 feet on the New Brunswick side is indicated by the existence of shells at that height and there is some evidence for 250 feet of emergence. On the Nova Scotia coast the highest figure is 137 feet which might suggest that there has been tilting up to the north-west. However, there are no dates for the 250-foot and 137-foot levels so they cannot be correlated.

Submergence of the Fundy coast is indicated in places by its outline, by the depth of river channels, by information from the continental shelf, by the existence of buried organic materials, and possibly by some historical evidence. The amount of submergence indicated by the first two lines of evidence is difficult to measure, but a figure of at least 150 feet is suggested although exactly when sea-level was 150 feet lower than at present is not known. In the case of buried organic materials, dating is easier, a figure of -39 feet for 4040 years ago having been recorded.

During the Recent Epoch there has been an interplay between isostatic forces, tending to produce emergence, and eustatic sea-level changes, tending to produce submergence. Soon after deglaciation, about 12,500 BP, Chignecto Isthmus was flooded, but isostatic recovery resulted in emergence of the Isthmus by 10,600 BP. Over the past 6,000 years the tidal range in the Bay of Fundy has been increasing which may account for the submergence of many of the trees and peat layers along the coast. However, tidal range may have reached its maximum, in which case high-tide level will decrease giving the impression of emergence.

The study here presented was commenced in 1965, and was aimed at a critical evaluation of existing knowledge and theory, with such additional work as time and circumstance might permit. New techniques rapidly became available as the work progressed, but with so large an area to cover it proved impossible to apply them as widely or thoroughly as the author would have liked. Many problems have thus had to be left incompletely investigated, but it is hoped that suggestions made will be useful to future workers in the area.

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CHAPTER I

INTRODUCTION

Location and Delimitation of the Area

The Bay of Fundy is an arm of the Atlantic Ocean that almost separates Nova Scotia from New Brunswick (Fig. 1). About 50 miles wide at its mouth, the bay narrows to the north-east and bifurcates at Cape Chignecto (Fig. 2). The southernmost branch includes the water bodies known as Minas Channel, Minas Basin and Cobequid Bay; the more northerly includes Chignecto Bay which forks at Cape Maringouin into Shepody Bay and Cumberland Basin. The latter is separated from Northumberland Strait by the low Chignecto Isthmus. The Bay has no clearly defined southern limit, but for the purpose of this study the name "Bay of Fundy" will be applied to the area of the sea to the north-east of a line joining West Quoddy Head in Maine, the southern tip of Brier Island, and Cape St. Mary in Nova Scotia. Grand Manan Island, at the mouth of the Bay of Fundy, will be referred to from time to time.

Clearly in any study of the evolution of a major landform it is necessary to go beyond the confines of that landform. This is particularly true in this case when dealing with the influence of glaciation during the Pleistocene and with the effect of eustatic changes of sea-level. The study is concerned principally with coastal morphology, but inland and submarine topography will also be discussed.

Map, Chart, and Air-Photo Coverage

Map and air-photo coverage of the coast is good. There is complete coverage by topographical maps of scale 1:250,000 and 1:50,000. Complete stereoscopic air-photo coverage of the coast at a scale of 1 inch to 1320 feet (1:15,840) is also available. Photographs of the New Brunswick side taken in 1967 can be obtained from the Department of Natural Resources, Fredericton and of the Nova Scotia coast from the National Air-Photo Library, Ottawa. Those of the Nova Scotia coast were taken in 1964 (northern part) and in 1967 (southern part). The northern part includes the coast of Cumberland Basin; Chignecto Bay; the north coast of Minas Channel, of Minas Basin, and of Cobequid Bay; and the south coast of Cobequid Bay and of Minas Basin as far west as the mouth of the Avon River. Included in the southern part is the south coast of Minas Basin west of the Avon River; the south coast of Minas Channel; the Fundy coast proper; Annapolis Basin and St. Marys Bay. The writer did not have easy access to the air-photographs until the study was nearly completed. Consequently, air-photo interpretation was done in retrospect rather than in preparation for fieldwork.

In addition to the maps and air photographs, several hydrographic charts of various scales are available from the Canadian Hydrographic Service, Ottawa. The most useful for general purposes are no. 4010, *Bay of Fundy (Inner Portion)*, scale 1:200,000 and no. 4011 *Approaches to Bay of Fundy*, scale 1:300,000. For more detailed studies there are large scale charts of parts of the Bay and of individual small harbours.

Finally, the Maritime Marshland Rehabilitation Administration (M.M.R.A.) which was concerned with the reclamation and preservation of tidal marshes in the Maritime Provinces surveyed some sections of the marshes at scales as large as 1:600. Some of the plans which have contours at one-foot intervals are particularly useful in plotting the exact location and determining the exact depth of submerged tree stumps.

Accessibility

A large part of the Bay of Fundy coast is accessible by gravel roads, but access is difficult: 1) from Salmon River, New Brunswick to Point Wolfe, New Brunswick; 2) from Spicer Cove, Nova Scotia to West Advocate, Nova Scotia (including Cape Chignecto); and 3) from Cape Blomidon, Nova Scotia to Cape Split, Nova Scotia (Fig. 3). Five large islands are found in the area of study. The two islands in Passamaquoddy Bay, Deer Island and Campobello Island, are relatively easy of access. Deer Island is connected by ferry to the New Brunswick mainland at Letite and a bridge joins Campobello Island to the United States' mainland at Lubec, Maine. Grand Manan Island can be reached by ferry from St. John, New Brunswick and from Blacks Harbour, New Brunswick, but it is more remote than the two islands mentioned previously and was not visited by the writer. On the Nova Scotia side, the two islands at the end of Digby Neck, Long Island and Brier Island, can also be reached by ferry.

The existence of good map and air-photo coverage is of considerable importance because, although it is easy to get to the shore from the landward side, except in the case of the three stretches mentioned above, progress along the shore is not always easy. In many cases it is possible to walk along the shore only at low-tide and as the tidal fluctuation is great this can be dangerous. To walk along the cliff top is also difficult because the region is, in most parts, heavily forested and progress is very slow. Approach from the sea would seem to be the answer to these problems, but the tides and frequent fogs make it imperative to hire not only a boat but also a man who knows the Bay well. While this might be feasible for a day or two in order to visit a single key area, prolonged use of such facilities would be expensive. Thus, since air-photos were not available for pre-fieldwork reconnaissance, it was necessary to select from the literature and the topographic maps those areas which seemed accessible and likely to yield the most useful information.

Previous Work

The Bay of Fundy has long attracted the attention of geomorphologists and geologists, but on the whole they have been concerned with parts of it only or with specific aspects of its evolution such as the effects of glaciation.

Interest in the area dates from the second half of the nineteenth century: Dawson's book, *Acadian Geology* (1855a), contains the first references to the geology and physiography of the area surrounding the Bay of Fundy; Dana (1863, p. 387) made what was possibly the first reference to "the Acadian basin"; Chalmers (1890, 1895) speculated on post-Tertiary changes of sea-level in the Bay; Matthew (1894) determined that the Saint John River had followed a variety of outlets during pre-glacial and post-glacial times; Bailey (1897) reviewed the geological history of the Bay of Fundy, starting with geosynclinal deposition and uplift during the Lower Palaeozic and finishing with continental glaciation and local glaciation during the Pleistocene; and Bailey (1898) described evidence of submergence and subsequent emergence of the Annapolis Valley - St. Marys Bay region.

Not until sixteen years later did Goldthwait (1914) renew interest in the area by estimating the marine limit at Saint John, Digby, and Truro. Powers (1916) made an extensive and detailed study of the rocks of the Acadian Triassic Basin. Goldthwait (1924) published a monograph on the physiography of Nova Scotia which included an examination of the coast and considered the question of changing sea -levels. He also observed glacial phenomena and theorized about the direction of ice movement over Nova Scotia. The following year, Johnson (1925) published his famous book, *The New England - Acadian Shoreline,* which contained a description of parts of the Fundy coast and a detailed study of the marshes at the head of the Bay. Both Goldthwait and Johnson referred to a peneplain, formed during the Cretaceous, which had been uplifted and tilted towards the south-east. As a result of a study of cross profiles of the Bay and of the Maine coast to the south, Johnson originated the idea that the Bay's north -west coast is a fault scarp and that the Bay of Fundy is a submerged lowland created by fluvial erosion of a Triassic fault trough. This idea was challenged by Shepard (1930) who proposed the alternative hypothesis that the steep slopes of the north-west coast are due to glaciation. The opposing faulting and glaciation hypotheses were the subject of a heated exchange between Koons (1941a, 1941b, 1942) and Shepard (1942).

Klein (1960) was principally concerned with the bedrock geology of the margin of the Bay of Fundy, but he arrived at some conclusions about the date of origin of the Bay and about the glacial/fault controversy. Swift and Borns (1967a, 1967b) studied a raised fluviomarine outwash terrace on the north side of Minas Basin, and Swift and Lyall (1967, 1968a, 1968b) obtained sub-bottom profiles in an attempt to determine the origin of the Bay.

Some workers have concentrated specifically on changes of sea -level: Lougee (1953, 1954), who was a proponent of the hinge-line theory of isostatic recovery, constructed several maps of his interpretation of land/sea/ice relationships during various stages of glaciation and deglaciation; Lyon and Harrison (1960) and Harrison and Lyon (1963), continuing work started much earlier (Lyon and Goldthwait, 1934), obtained a chronology for movements of sea-level in the Bay of Fundy during the period 4500-3000 BP; and Grant (1970) conducted a thorough study of the Atlantic and Fundy coasts of Nova Scotia to determine comparative rates of change of sea-level.

In 1967, the Geological Survey of Canada started a programme aimed at detailed mapping of the Pleistocene deposits between Saint John, New Brunswick, and the United States' border. As a result, Gadd (1968, 1969, 1970a) arrived at some conclusions about the glacial and post-glacial history of this part of the coast, and obtained information about sea-level changes.

Geological Survey of Canada memoirs published on a number of areas around the Bay usually contain a short description of the physiography of the area in question with occasional references to changes in sea -level. Memoirs have been published by Wright (1922), Bell (1929), Alcock (1938), Weeks (1948), Hughes (1957), Stevenson (1958, 1959), Crosby (1962), and Taylor (1969).

Several doctoral and masters theses contain at least some reference to the physiography of the Fundy coast and to evidence of changes in sea-level: Kiewiet de Jonge (1951), Crosby (1951), MacNeil (1951a), Purdy (1951), Dunlop (1952), Swayne (1952), Hickox (1958), Hudgins (1960), and Melvin (1966). With the exception of those by Kiewiet de Jonge and Melvin, they are all concerned with the Nova Scotia coast.

In general, the Bay of Fundy and the adjacent areas of New Brunswick and Nova Scotia are receiving increased attention from geomorphologists and Quaternary geologists. Workers based at the Atlantic Oceanographic Laboratory, Dartmouth, Nova Scotia, are obtaining information about submarine conditions and the Geological Survey of Canada is showing increasing interest in the area. In addition, people based at research institutes and universities in the north-east of the U.S.A. are finding the area a fruitful one for research. It is to be expected, therefore, that there will be a steady increase in knowledge of the area and that some long-standing problems will be solved, or as nearly solved as anything is in geomorphology.

In any study covering a large area and extending over a long period of time, the work is never complete; new information comes to light and other workers publish articles containing new material which may support or contradict the writer's opinions. In order to overcome this problem, articles published and information obtained before July 1, 1970, the date at which writing started, are referred to but material becoming available after that date has not been considered.

Purpose and Aims

As has been shown, there are several studies of parts of the Bay of Fundy and its coast and also studies dealing with specific aspects of its morphology, but there is no comprehensive account of the evolution of the Bay as a whole. Perhaps the accounts of Bailey (1897) and Swift and Lyall (1968b) are the most comprehensive but the latter concentrated heavily on the glacial/fault controversy. While the writer intended to try to produce a comprehensive account, this proved to be a major undertaking and it soon became obvious that certain aspects would have to be omitted or treated in little detail. As always, the more reading and research that was done the more complex the problem became and the greater the number of unanswered questions. This study should, therefore, be regarded as a detailed reconnaissance rather than as a complete account of every aspect of the morphology and the evolution of the Bay of Fundy, although it is an attempt to synthetize previous work and the writer's own work on the topic.

When the writer began this study he had seven aims in mind:

(1) to write an account of the present physiography of the Fundy coast and, as far as possible, to describe the submarine conditions of the Bay of Fundy and adjacent sea areas;

(2) to test the validity of some long-cherished ideas about uplifted peneplains in New Brunswick and Nova Scotia;

(3) to examine the evidence for glaciation of the area and to establish the relationship between glaciation, tidal fluctuations and the interpretation of emerged and submerged features formed at or close to sea-level;

(4) to investigate the use of pebble-shape analysis as a means of distinguishing gravel beds of various origins;

(5) to locate and assess the significance of evidence of higher sea -levels during the Quaternary;

(6) to locate and assess the significance of evidence of lower sea -levels during the Quaternary;

(7) to establish a chronology of Quaternary events.

The structure of the dissertation closely follows this list except that the work on pebble-shape analysis has been omitted. Chapter 2 is purely descriptive and aims at acquainting the reader with the present nature of the Bay with respect to geology, coastal morphology, submarine topography and tides. Chapter 3 contains an attempt to summarize what is known about the Bay and adjacent regions as they evolved during the great span of pre-Pleistocene time. Chapter 4, which deals with the Pleistocene, is a summary of what is known about the glaciation and deglaciation of New Brunswick and Nova Scotia and of adajcent areas now below sea-level. Chapter 5 is concerned with changes of sea-level during the Quaternary. The first part of the chapter is devoted to discussion of some of the problems encountered when interpreting coastal features supposedly of Quaternary origin. The difficulties having been described, an attempt is made to analyse the evidence both for higher sea-level during the Quaternary and for lower sea-level during that period. The chapter finishes with a tentative chronology of events for the Quaternary. Finally, in Chapter 6, the writer's conclusions are presented and some suggestions are made for future research in the area.

Time-Table and Method of Study

The writer first considered undertaking this project in the spring of 1964. During July and August of that year a reconnaissance survey of the Fundy coast was carried out and interesting localities were noted and marked for detailed study at a later date. A few stretches of the coast were described in as much detail as purely visual methods and map reading permitted.

July 15 - August 22, 1965, fieldwork was undertaken in New Brunswick and Nova Scotia, the work completed falling into two categories:

(1) Investigation of the marshes at the head of the Bay of Fundy.--A hammer seismograph was used to determine the depth of silt at a number of sites; detailed sections of parts of the marshes were obtained by augering; and peat samples and samples of wood from buried trees were taken for radiocarbon dating, two of the wood samples being subsequently dated (I-4312: I-4313).

(2) Study of the coast between St. Stephen and St. George, New Brunswick.--The distribution of glacial deposits was determined on a reconnaissance basis. Shells collected from one site near St. Andrews were submitted to the Geological Survey of Canada for radiocarbon dating. The quantity of shell material was insufficient

for dating, but in 1967 N. R. Gadd of the Geological Survey of Canada visited the site and was able to collect enough material for a date to be obtained (GSC-795). Pebble samples, collected from deposits thought to be of glacial, glaciofluvial and marine origin, were subjected to shape analysis to determine whether there is any clear relationship between shape and mode of origin. However, it was decided that not enough control had been established from samples known to have been deposited under specific conditions; thus the writer could not use the results to speculate about deposits of unknown origin. The results are, therefore, not presented in this thesis.

During the period June 1 - August 17, 1966, emphasis was placed on the identification, location and interpretation of indications of higher past sea-levels. Although some work was done along the Nova Scotia coast of Chignecto Bay and along the north coast of Minas Channel, Minas Basin and Cobequid Bay, most of the time was spent on the New Brunswick coast. Of several shell samples collected, one was dated by the Geological Survey of Canada (GSC-965).

From May 22 - June 21, 1967, a quick survey of the south coast of Cobequid Bay, Minas Basin, Minas Channel, and the Nova Scotia coast of the Bay of Fundy proper was carried out. Some indications of former high sea-levels were located and one shell sample which was collected has since been dated by the Geological Survey of Canada (GSC-887). Of the wood samples collected from a "submerged forest" near the junction of the Kennetcook and Avon rivers one has been dated (I-4314). In addition an attempt was made at deep boring of the Tantramar Marshes at the head of

Cumberland Basin to try to locate and sample peat bands below present sea-level.

Analysis and interpretation of the results obtained by fieldwork and of information obtained by reading the relevant literature have taken place sporadically from the summer of 1967 to the date of thesis submission. After fieldwork was completed, the writer was able to obtain, with the aid of a grant from the Defence Research Board of Canada, complete stereo-photo coverage of the Fundy coast. Since the photographs were obtained, an air-photo interpretation of the coast has been carried out using observations made in the field during the summers of 1964-1967 as ground control. One of the results is a morphological map of the Fundy coast (Fig. 4).

CHAPTER 2

DESCRIPTION

Physiographic Divisions of New Brunswick and Nova Scotia

Although this work is concerned primarily with the Bay of Fundy, it is impossible to describe and attempt to explain the evolution of this landform without reference to the general physiography of the area in which it is located. A concise summary of the physiography is given by Weeks *in* C. H. Stockwell (1957) and the following description is, with minor modifications, taken from his account.

Mainland Nova Scotia, excluding Cape Breton Island, can be divided into two highlands, an upland, and two lowlands (Fig. 5). The highlands are the Cobequid Mountains extending parallel to Cobequid Bay from Cape Chignecto east to the vicinity of Pictou, and the Antigonish Highlands extending south and west from Cape George. The Cobequid Mountains, with a gently rolling surface about 10 miles wide, reach elevations of over 900 feet. The Antigonish Highlands, being smaller, are more dissected than the Cobequid Mountains, but have the same type of surface and reach the same elevations.

The uplands of Nova Scotia include North Mountain and the Atlantic Uplands. North Mountain is a narrow, flat-topped belt, averaging about 550 feet high and extending along the south-east side of the Bay of Fundy from Cape Blomidon in Minas Basin south-west for 120 miles to

Brier Island. The Atlantic Uplands of Nova Scotia lie along the entire coast from Cape Sable to Cape Canso (and are continued onto Cape Breton Island). The surface of this upland is supposed to represent a Cretaceous peneplain that slopes to the south-east from elevations of about 600 feet in the north-west where it is referred to as South Mountain, to sea-level along the Atlantic shore.

The principal lowland areas are: the Cumberland Lowland lying on the north and east sides of the Cobequid Mountains and extending north to Northumberland Strait; the Minas lowland surrounding Minas Basin and Cobequid Bay on all sides; and the Annapolis Valley which lies between North Mountain and South Mountain.

New Brunswick falls naturally into one major highland and one major lowland. The highland, the New Brunswick Highland, is U-shaped with the west arm covering much of the central part of the province, the curved part occupying the southern part, and the east arm bordering the Bay of Fundy. The southern part of the "U" is crossed by the Saint John River which divides the highland into three parts; the Miramichi Highlands north-east of the river, the St. Croix Highlands west of the river, and the Caledonian Highlands east of the river. The flat surfaces of the highland are supposed to represent the south-east-sloping Cretaceous peneplain, the north-west parts being higher than those bordering the Bay of Fundy; for example, the Miramichi Highlands reach elevations greater than 2000 feet whereas the St. Croix and Caledonian Highlands reach only about 1000 feet.

The New Brunswick Lowland is an area almost enclosed by the U-shaped New Brunswick Highlands. This lowland, Prince Edward Island,

and the Magdalen Islands comprise a single physiographic unit. Whereas the New Brunswick Highlands are underlain by sedimentary, volcanic and other igenous rocks, the lowland areas are underlain by soft, Carboniferous rocks.

The Coast

Terminology

Some ambiguity exists with respect to the terminology of coastal and shore features. Therefore, in an attempt to avoid confusion, the terms that are most commonly used in the text are defined in this section. However, as Arnold (1937, p. 180) pointed out: "Definition is ordinarily supposed to produce clarity in thinking. It is not generally recognised that the more we define our terms the less descriptive they become and the more difficulty we have in using them." In the case of the Bay of Fundy it is even difficult to define "river" and "sea" because the great tidal fluctuation of the "sea" affects the mouths and, in some cases, considerable stretches of the "rivers". Thus, except where a waterfall marks the landward limit of tidal penetration, the change in terminology from "river" to "sea" is usually based on some arbitrary consideration. Use of the term "estuary" which has been defined as "that part of the lower river course that is affected by the mixing of salt water and fresh" (Fairbridge 1968, p. 326) might seem to be the answer to this problem. However, in the case of the small "rivers" emptying into Cumberland Basin and Shepody Bay, the mixing of fresh and salt water takes place along practically their entire length.

Because of the difficulty of definition and the restrictions resulting from rigid definitions, the number of terms defined is kept to a minimum.

Shore: "The zone extending from the low-tide to the landward limit of effective wave action." (Thornbury 1954, p. 433) It is, in fact, a broad zone across which the land/water line moves. In the case of the Bay of Fundy with its large tidal range, the zone may be several miles wide.

Shoreline: "The position of the water level at any given time and [this] varies between the low-tide shoreline and high-tide shoreline." (Thornbury, 1954, p. 433) The two terms "shore" and "shoreline" are frequently used interchangeably and it is often difficult to make a clear distinction between the two.

Backshore: "The zone above the limit of the swash of normal high spring tide and is, therefore, only exceptionally under the direct influence of the waves. On a rocky coast it will include the cliffs and on a low coast it may consist of sand dunes or mature salt marsh." (King 1959, p. 48)

Foreshore: "That part of the beach [or wave-cut platform] which is regularly covered and uncovered by the tide." (King 1959, p. 48) The words in brackets have been added by the writer because it is possible that the tide will regularly cover and uncover areas which have no beach as defined later. As with the more general term "shore", in the case of the Bay of Fundy the zone may be miles wide.

Offshore: "The 'offshore' zone extends from the uppermost point always covered by water to a depth at which substantial movement of beach material ceases." (King 1959, p. 48)

Coast: "A zone of somewhat indeterminate width which extends landward from a shore or shoreline. The boundary between it and the shore is the coastline." (Thornbury 1954, p. 433) Although there is a distinction between "shore" and "coast", it is frequently convenient and certainly less clumsy to refer to the general area of land/sea contact as "the coast" rather than as "the shore and coast". The word "coast" will on occasion be used in this way in the text.

Sea cliff: "A knick or scarp resulting from wave erosion." (Thornbury 1954, p. 433) When it is obvious that a coastal region is referred to, the word "sea" is omitted without any loss of clarity.

Wave-cut Platform: "The rock shelf that is produced by the combined action of the direct attack on the cliff base, the to-and-fro motion on the wave base, and the undertow." (Fairbridge 1968, p. 859) The platform extends seaward from the base of the cliff.

Beach: "The zone of unconsolidated material extending landward from the mean low water line to the place where there is a change in material or physiographic form as, for example, the zone of permanent vegetation, or a zone of dunes, or a sea cliff. The upper limit of the beach usually marks the effective limit of storm waves." (Shepard 1963a, p. 169)

Spit: "A ridge or embankment of sediment attached to the land at one end and terminating in open water at the other." (Evans 1942, p. 846)

Bar: "A completed or extended spit which encloses, or nearly encloses, a portion of the water body into which it extends." (Evans 1942, p. 846)

Descriptions of Individual Sections of the Coast

Geology, exposure to erosion, tidal conditions, coastal deposition, and human interference have resulted in a variety of coastal types around the Bay of Fundy. The rocks exposed along the Fundy coast range in age from Precambrian to Recent (Fig. 6). This does not mean, however, that the whole geological column is represented. No rocks of between Triassic and Pleistocene age are exposed along the coast. In order to avoid repetition and possible confusion, a summary of the stratigraphy of the Bay of Fundy area (from the Ordovician onwards) is included (Fig. 7), together with a more detailed stratigraphic chart of the Triassic formations of the Maritime Provinces (Fig. 8).

Grand Manan Island

The influence of geology on the form of the coast is nowhere better illustrated than on Grand Manan Island. The western part of the island is composed of Triassic basalt whereas the eastern part is composed of a variety of lithological types of Precambrian age. Along the straight west coast the basalt forms cliffs up to 400 feet high. In marked contrast, the east coast is low-lying with few cliffs, many inlets, and numerous small islands. The most notable depositional landform is a bay-mouth bar at Dark Harbour on the west coast.

Cobscook Bay - Passamaquoddy Bay

The unifying characteristic of this area is the intricate plan of the coastline. Cobscook Bay, in particular, is a maze of inlets and peninsulas, and it is clear that the coastal outline is determined by the structure. Johnson (1925, p. 15) referred to this area as a "young shoreline of submergence ... showing concentrically curving bays, peninsulas and islands due to partial submergence of curving valleys and ridges developed by normal erosion on folded rocks of varying resistance."

There is no such obvious structural influence in the case of Passamaquoddy Bay, although it is characterised by numerous islands, including Campobello Island and Deer Island, and many peninsulas. Campobello Island, Deer Island and many of the adjacent smaller islands are underlain by an assemblage of sedimentary and volcanic rocks of Silurian age. Where these rocks reach the sea, the seaward slope is steep but there is usually no cliff.

The same is true of much of the mainland coast of Passamaquoddy Bay; for example, the tidal part of the Magaguadavic River below the waterfall at St. George flows in a steep-sided valley but erosion at the present sea-level has had little effect, except at Midjik Bluff on the south side of the river's mouth. Here, the Perry Formation (Upper Devonian) consisting of red conglomerates, sandstones, and shales which are much more easily eroded than the Silurian rocks, is exposed.

Pleistocene deposits of various origins cover much of the area. In places they are thick and where they reach the coast they are easily eroded to form cliffs; for example, at Sand Point and at the southern point of Navy Island.

Some major depositional landforms exist:

(1) In Maine, West Quoddy Head is joined to the mainland by a short tombolo and a mile further north a spit extends southward from South Lubec.

(2) On Campobello Island are two excellent examples of bars, the more impressive of the two being at Herring Bay on the east side of the island. The bar, which is over a mile long, extends northward from near Dinner Head and almost completely separates Lake Glensevern from the sea. The second example is a mid-bay bar at Mill Cove. It is much shorter than Herring Bay Beach but extends nearly across the Cove, leaving only a small outlet to the sea.

(3) Minister Island is joined to the St. Andrews peninsula by a bar which is exposed at low-tide but covered at high-tide.

(4) A sand bar extends from the north-east side of Navy Island towards the mainland at St. Andrews.

With the exception of a small outcrop of the Perry Formation at Blacks Harbour, the rocks which are of Precambrian and Silurian age are hard and resistant to erosion so that at most places cliffs are low or absent. Johnson (1925, p. 162) described the "low cliffs at the base of higher hill slopes" at Letang Harbour as being "characteristic of an early stage of development of a young shoreline of submergence."

Coastal deposition along this stretch of coast is chiefly represented by tidal mud-flats, for example at Pocologan Harbour, but at New River Harbour is found one of the few good sand beaches along the Fundy coast.

<u>Lepreau Harbour to Dipper Harbour</u>

This relatively short stretch of coast is distinguished separately because at Lepreau Harbour and between Little Lepreau Basin and Dipper

Harbour are outliers of Triassic rock - the Lepreau Formation (Fig. 8). The most westerly of several such outliers found along the New Brunswick coast, they consist predominantly of red clastic rocks - conglomerates and sandstones. At Lepreau Harbour the Triassic rocks occupy a narrow belt trending north-east to south-west on either side of the estuary. The rocks are relatively easily eroded and between Little Lepreau Basin and Dipper Harbour there are some steep cliffs. These are not high (less than 50 feet), but in places they are nearly vertical and at their base there is a wide wave-cut platform particularly at Maces Bay (Fig. 9).

Dipper Harbour to Lorneville Point

This section of the coast is diverse both in outline and in profile, consisting of a series of headlands and deep inlets, and one stretch of straight coast. The diversity is a reflection of a great range of rock types including igneous rocks, both intrusive and extrusive, and several kinds of sedimentary rock. It is increased by the accumulation, in favourable localities, of large areas of tidal sand-, silt-, and mud-flats.

From Dipper Harbour to Little Dipper Harbour the coastline is irregular and jagged but cliffs are, on the whole, low or absent. Both Dipper Harbour and Little Dipper Harbour, wide inlets of the sea, contain extensive areas of sand and mud which are uncovered at low-tide and at Little Dipper Harbour a spit has developed, nearly separating the innermost part of the inlet from the open sea.

From Little Dipper Harbour to Western Head the coast consists of a series of inlets: Chance Harbour, Harbour by Chance, Little Musquash

Cove, and Gooseberry Cove. The vertical profile of this section of the coast varies. At the backs and sides of the inlets there is a relatively gentle slope down to the sea, but on the headlands there are some steep cliffs; for example, between Little Musquash Cove and Gooseberry Cove they are 100 feet high.

The lower reaches of Musquash River are tidal downstream from the small settlement of Musquash. Seaward from the settlement, the valley of the Musquash River consists of three broad areas of tidal mud-flats and marsh separated from each other by constrictions. The innermost of the three is mainly marsh-covered and is flooded only at very high tides, whereas the second is more susceptible to flooding and has a lower percentage of its area covered by marsh. At Five Fathom Cove the river is channelled through a narrow gap and into Musquash Harbour where extensive areas of tidal mud- and sand-flats exist along the shore. Marsh, however, is less common than in the case of the two innermost tidal flats. The Musquash River enters the Bay of Fundy through a gap between Western Head and Musquash Head.

The Musquash tidal flats are the first good example of the tidal mud-flats and marshes which are such a distinctive feature of the Fundy coast, especially in its upper parts.

Between Musquash Head and Lorneville Point the strike of the rocks is south-west/north-east, parallel to the direction of the coast. The cliffs between Tiner Point and Negro Head are steep but have no obvious wave-cut platform at their base. It is worth noting at this point that, compared with the coasts described so far, this section is extremely straight and is about parallel to a stretch of coast further to the north-east between Quaco and Salisbury Bay which Johnson (1925, pp. 39-40) described as a "fault-line-scarp shoreline".

Lorneville Point to Black Point

A complex stretch of coast, it is significantly different from any so far described in that in places there are high cliffs cut into Pleistocene deposits. In fact this area, which includes the mouth of the Saint John River and Saint John Harbour, is underlain by rocks ranging in age from Precambrian to Recent.

Lorneville Harbour and Manawagonish Cove constitute an extensive area of tidal sand- and mud-flats up to a mile wide. Manawagonish Cove is backed on its west and north-west by a coast of granite, a sharp contrast to the tidal flats. At the north end of the Cove, however, marshy land is drained by Manawagonish Creek. Taylor Peninsula, an island of bedrock, has been joined to the mainland by deposition. East of the peninsula, almost to Sheldon Point, the coast is formed by Pleistocene deposits consisting in the main of marine clays and glaciofluvial deposits. The red marine clays are easily eroded and have been cut back by the sea to form steep cliffs (Fig. 10) although in places slumping has reduced their gradient. From Sheldon Point to Sand Cove, bedrock outcrops along the coast, but from the latter to Negrotown Point the coast is mainly developed on Pleistocene deposits. The red clay mentioned earlier is found along this stretch of coast and in places contains some very large boulders. Where the cliffs are rapidly eroded by the sea, there is much slumping with the large boulders being gradually isolated from the clay and left on the beach.

At the mouth of the Saint John River and in Saint John Harbour the rocks exposed are of varied lithology and give rise to a coast that is intricate in outline. The best known feature of this part of the coast is the falls, or more correctly the rapids, at the mouth of the river which flows through a narrow gorge before entering Saint John Harbour. If the present situation is viewed without any consideration of the geomorphological history of the Saint John area, it is difficult to understand why the river cuts a deep gorge through very resistant Precambrian rocks when there are much easier routes to take.

The coast on the east side of Saint John Harbour is mainly developed in surficial deposits although there are some bedrock outcrops. There are some similarities between this side of the Harbour and the west side, the main one being the exposure along the cliffs of red clay. At places such as Red Head where the clay is being eroded by the sea, the lower parts of the cliffs are almost vertical; but in other places the clay, which is overlain by sand and gravel, has slumped badly to give irregular cliffs. As with the red clay at Negrotown Point, large boulders are embedded in it and are left on the beach as the clay is gradually removed. Sand- and mud-flats are extensive between Courtenay Bay and Cranberry Point. A mile north of Black Point, bedrock is exposed and from this point eastward the nature of the coast changes abruptly.

Black Point to Quaco Head

The coast is, for the most part, developed on rocks of Mississippian or Pennsylvanian age, but two small outliers of the Quaco Formation of Triassic age (Fig. 8) occur at Robinson Cove and Honeycomb Point

(Alcock, 1938). Also, at Mispec Park, red clay is exposed at the base of irregular wood-covered cliffs. This clay, similar to that found east and west of Saint John Harbour, represents the most easterly exposure of it found by the writer.

The cliffs along parts of this coast are impressive. At Cape Spencer they are in the order of 150 feet high and at their base enormous boulders have accumulated. Cape Spencer is one of the most exposed parts of the Fundy coast with extensive fetch to the south-west. Consequently, coastal erosion is rapid even though the rocks exposed are resistant. North-east of Cape Spencer, as far as West Bay, the cliffs are steep and there are no inlets; therefore access is difficult. The coast has a south-west/north-east trend and is relatively regular, being similar in many ways to the area between Tiner Point and Negro Head. Some prominent stacks have been formed; for example, on the west side of Robinson Cove, at East Red Head, at Rogers Head, and at Quaco Head.

Depositional landforms are represented by several spits and bars built across the mouths of small streams: Emerson Creek; Gardner Creek; a small stream that has been damned back to form Duck Pond; Tynemouth Creek (Fig. 11); and a small stream that has been damned to form Griffin Pond.

<u>Quaco Head to Macomber Point</u>

This short stretch of coast is described separately because it is underlain by rocks of the Quaco and Echo Cove Formations of Triassic age (Fig. 8). The Quaco Formation, consisting of a thick boulder conglomerate, is overlain by the red beds (sandstones and shales) of the Echo Cove Formation (Klein, 1962). The former gives rise to some prominent cliffs, most notably to the east of the settlement of St. Martins, and in some places lines of weakness have led to the formation of shallow caves (Whittle, 1891).

Behind Quaco Bay, coastal deposition has resulted in an area of marsh nearly a mile wide at its maximum, and at the south end of Quaco Bay the Mosher River is diverted southward by a spit growing across its mouth.

Macomber Point to Point Wolfe

This section of coast is straight and is part of what Johnson (1925, pp.39-40) referred to as a "fault-line-scarp shoreline". Backed by the Caledonian Highlands (Fig. 5) which are composed of resistant Precambrian rocks, it is a steep coast making access to the shore difficult. At places, for example to the south of the mouth of Fuller Brook, the cliffs are nearly vertical. Many small streams flow to the sea from the Caledonian Highlands and all have steep long and cross profiles.

In addition to the results of coastal erosion there are some good examples of coastal deposition. Berry Beach is a fine example of a shingle spit. Further east, Melvin Beach and Long Beach are sand beaches that have accumulated in alcoves of the coast. Martin Head is an island that has been tied to the mainland by a shingle bar that has grown eastward from the mainland. A spit extending southward from the north side of the mouth of Quiddy River nearly blocks its exit to the sea (Fig. 12).

Point Wolfe to Dennis Beach

This part of the coast differs from the previous section in that it is developed on sedimentary rocks of Carboniferous age, including conglomerates, sandstones and shales. The coast is usually steep and at the base of the steep slope there are large boulders (Fig. 13). Although it is more correct in most cases to talk of a seaward slope rather than of a cliff, some cliffs do exist, for example at Matthews Head, Joel Head, and Owl Head (Fig. 14). The two main rivers that reach the sea in this section, the Upper Salmon (Alma) River and the Point Wolfe River, have steep-sided valleys along their entire courses.

The coastal depositional landforms include a mid-bay bar across the mouth of Point Wolfe River and a small tidal delta formed by the Upper Salmon (Alma) River described in detail by Ali (1964) and Ali and Lamming (1966).

Dennis Beach to Harvey Bank (Fig. 15)

This is a distinctive stretch of coast, its nature being determined very largely by geological structure. Rocks of Mississippian and Pennsylvanian age strike in a north-east/south-west direction and outcrop as narrow belts of differing hardness. A small outlier of Triassic rock lies unconformably on the older rocks in the south. The Carboniferous rocks produce a series of parallel north-east/south-west trending ridges separated by marshy valleys occupied by tidal streams, the largest of which is the Shepody River which flows to the north-east.[1] The marshes present a marked contrast to the ridges on either side (Fig. 16).

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[1]The Shepody River is no longer tidal because a control dam has been built near its mouth by the M.M.R.A.

From the western end of Dennis Beach to Cape Enragé the general trend of the coast is at right angles to the strike of the Carboniferous rocks and consequently to the ridges and valleys. Dennis Beach, an almost straight stretch of shore, is over a mile long. At each end of it, the cliffs are composed of Triassic red sandstone and at Red Head at the eastern end, there are some prominent stacks (Fig. 17). Between the sandstone outcrops, the cliff face is composed of slumped sand and gravel which probably cover the Triassic bedrock.

East of Waterside, another stretch of straight coast some two miles long differs from Dennis Beach in that it is not backed by cliffs. Its name, Anderson Hollow, gives some clue as to its origin: clearly it was an inlet of the sea at one stage but vast quantities of shingle have been deposited across the mouth of the inlet so that Newfoundland Creek, which at one point is only a few hundred yards from the sea, is now diverted a mile to the south-east. Behind the shingle ridge is an extensive marsh area.

East of Anderson Hollow, the coast changes abruptly and two prominent ridges, one of which terminates at Cape Enragé, project southward into Chignecto Bay. Between the two, and in great contrast, lies Barn Marsh. A beach, built up at its southern end and joining the two ridges on either side, is composed of shingle and boulders, many of which are disc-shaped, together with an enormous amount of drift wood (Fig. 18).

From Cape Enragé northward the coast is regular and is almost parallel to the rock outcrops except where the sea has cut through the Boss Point Formation to erode the less resistant Enragé Formation. Between

Bray Beach and Two Rivers Inlet the coast is set back a quarter of a mile from the general trend to north and south and in this section sea water occupies the lowland, occupied by Barn Marsh to the south and Horton Marsh to the north.

For three miles north of Two Rivers Inlet, the coast is regular in outline, but north of this the strike of the rocks swings round to be more nearly west-south-west/east-north-east. As at Two Rivers Inlet, the sea has cut through the Boss Point Formation into the Enragé Formation. In this case, a shallow bay has been formed and is occupied by New Horton Flats. To the north and south of the bay, the Boss Point Formation produces headlands.

Shepody Bay: Harvey Bank to Cape Maringouin

The main characteristic of this stretch of the coast is the occurrence of large areas of tidal flats and marsh. This is certainly one of the most obvious features of the whole of the north-eastern and eastern part of the Bay of Fundy (Shepody Bay, Cumberland Basin, Minas Basin and Cobequid Bay). Vast areas of tidal marsh have been reclaimed, a process started by the earliest Acadian settlers who developed a system of dykes along the shore and along the rivers. There are, of course, still some "natural" marshes along the coast and as the tidal range increases towards the extremities of the Bay so does the extent of tidal marsh and of tidal sand-, silt-, and mud-flats. At Grande Anse, a shallow embayment on the east side of Shepody Bay, the shoreline retreats nearly two miles at low-tide to expose a vast expanse of sand and mud.

North of the mouth of Shepody River and south of Hopewell Hill, a traverse taken in a north-west/south-east direction would be typical of many areas in the northern and eastern extremities of the Bay of Fundy (Fig. 19). Route 14 at Hopewell Hill is built on bedrock. To the south-east the land slopes down to flat, reclaimed marshland protected by a dyke. South-east of the dyke there is a stretch of marshland which, because unprotected, is susceptible to occasional flooding. Further south-east is an area of tidal mud-flats at least a mile wide at this point.

However, at Hopewell Cape this arrangement does not occur. Here a seaward projection of bedrock, composed mainly of a coarse conglomerate of Pennsylvanian age, is exposed along the coast for a distance of about a mile. Erosion by the sea, working along lines of weakness, has led to the formation of some peculiarly-shaped stacks (Fig. 20). A miniature version of the same thing is found east of Edgett's Landing.

On the east side of Shepody Bay, bedrock coasts are more common. In several parts the coast is cliffed, particularly at Dorchester Cape where the cliffs are composed of red sandstone and conglomerate of the Boss Point Formation. Further south, behind Grande Anse, the cliffs are composed of shale below with a capping of surficial material. Further south still, as far as Cape Maringouin, bedrock outcrops of Carboniferous age form a series of parallel bands at right angles to the coastline, but the cliff face is composed almost entirely of surficial materials.

<u>Cumberland Basin: Cape Maringouin to Boss Point</u>

This area is the coast and shore of Cumberland Basin. Here the word "coast" and "shore" have to be interpreted liberally because the land is low-lying and the area influenced by tide-water extends far inland. North of Cumberland Basin are Tantramar Marsh and Aulac Marsh, and to the east is Amherst Marsh. Projecting above the marshes are two low but prominent ridges, Fort Cumberland Ridge and Fort Lawrence Ridge, composed of sandstone.

Several rivers and streams enter Cumberland Basin: with the exception of the Tantramar and Nappan, which have had dams built near their mouths to prevent the entry of salt water at high-tide, all are tidal. In the case of Aulac River, Missaguash River, La Planche River, Maccan River, and River Hébert, a great contrast is evident between the full river channel at high-tide and the almost empty channel at low tide (Fig. 21).

The inner parts of the marshes have been reclaimed for so long and are so well-drained that to use the word "marsh" gives a false impression. However, closer to Cumberland Basin the marshes have not been reclaimed and are occasionally covered by the tide.

Cumberland Basin, like the rivers that flow into it, varies in its appearence according to the state of the tide. At high-tide it is an area of muddy, reddish water, but at low-tide it is a vast expanse of red mud through the middle of which wanders a narrow water channel.

Cliffs are rare around Cumberland Basin but near its mouth there are some, for example at Pecks Point. Although tidal sand- and mud-flats are extensive, the wave-cut platform is occasionally exposed offshore from promontories as at Pecks Point, Black Point, Wood Point, Minudie Point, and Boss Point.

Boss Point to Squally Point

This section of the coast is developed on rocks of Pennsylvanian age. Although there is a capping of surficial material in places, in most cases the cliffs are cut into bedrock. At many points the cliffs are high and steep; for example, north-west of Joggins they are 100 feet high and almost vertical: and at other places there are cliffs over 200 feet in height.

Further south, near Pudsey Point, cliffs are cut into a conglomerate much like that at Hopewell Cape in New Brunswick, and at Spicer Cove the steep cliffs are cut into soft red sandstone. One remarkable feature is that, although these cliffs are clearly being eroded rapidly, there is no accumulation of boulders on the beach. Probably the rock is so soft that it is readily converted into sand.

Along nearly all of this stretch of coast there is a well-developed wave-cut platform, in some places partially covered by gravel, sand and mud, but in others with no beach covering (Fig. 22). It was observed at several places, notably at Joggins and near the mouth of Flat Brook, that at high-tide the sea reaches well above the base of the cliffs.

In addition to the erosional features, there are some conspicuous depositional features. The area between Apple Head and Cape Capstan was probably an island in the past for to the east it is joined to the mainland by a low-lying area occupied by Polly Bog. The mouth of Apple River exhibits many of the characteristics of the tidal rivers at the head of Cumberland Basin—extensive areas of dyked marsh and, outside the dyked areas, tidal sand- and mud-flats. North of West Apple River, a hooked spit projects northward across the mouth of the river.

Squally Point to McGahey Brook

This section of coast is probably the least accessible of any around the Bay of Fundy. The rocks making up the coast are granite and allied igneous rocks whose geological age is not known. The result of the outcrop of these resistant rocks is an extremely steep, rugged coast with many stacks, especially along the north/south section (Fig. 4). Streams flowing to the coast are short, flow rapidly and have created steep-sided valleys, especially those flowing southward to Advocate Bay. The coast between Cape Chignecto and the mouth of McGahey Brook follows the line of the Cobequid Fault, the downthrown side being to the south (Cameron, 1949). It is remarkably straight and steep - at one place a height of 750 feet is reached within a quarter mile of the shore. This rise is in part sea cliff and in part seaward slope.

McGahey Brook to Wards Brook

This section of coast is less uniform, a greater variety of rock types being exposed. The Cobequid Fault runs inland between McGahey Brook and Parrsborough Shore and then along the coast again as far as Wards Brook. In the land area to the south of the fault line, rocks of Pennsylvanian and Triassic age are exposed. The rocks of Triassic age consist of two groups of vastly different lithology and resistance to erosion. The older Triassic rocks, belonging to the Annapolis Formation (Fig. 8), consist of conglomerates, sandstones and shales which are, in most cases, easily eroded. The younger Triassic rocks are extrusive igneous rocks - mainly basalt - and are much more resistant to erosion.

A belt of basalt, extending from Cape D'Or to Cape Spencer, forms highland reaching over 500 feet; where it reaches the coast there are impressive steep cliffs over 250 feet high in places, for example north of Cape D'Or (Fig. 23). It is noteworthy that this cliff line is at right angles to the maximum fetch from the south-west along the axis of the Bay of Fundy. The basalt between Cape D'Or and Cape Spencer is an extension of the basalt that forms North Mountain, terminating at Cape Split, south of Minas Channel. Where the Annapolis Formation, which underlies a belt of country to the north of the basalt, reaches the sea it produces a low-lying coast in marked contrast to that developed on the basalt.

At Advocate Harbour a long spit extends south eastward from West Advocate and nearly joins up with another one which extends northward from the basaltic highland (Fig. 24). Behind the two spits are tidal mud-flats and north of the mud-flats is an area of dyked marsh.

From about one mile south of Spencers Island to a point just east of Parrsborough Shore the coast, which is smooth in outline and which has occasional steep cliffs, is cut into rocks of Pennsylvanian age. Between Parrsborough Shore and Wards Brook, where it follows the Cobequid Fault and is developed on rocks of Silurian age, the coast is straight and steep, although the cliffs are not vertical. In places they are extremely irregular in profile due to variations in lithology, and there are one or two prominent stacks, for example east of Brookville. As the slope down to the coast is steep, the streams that flow to Greville Bay are short, fast-flowing and occupy steep-sided valleys.

Wards Brook to Parrsboro

In this section the Cobequid Fault is set back from the coast and forms the steep, south-facing slope of the Cobequid Mountains. The bedrock is mainly Pennsylvanian in age, but at Cape Sharp and Partridge

Island the rock exposed is Triassic basalt. Between the coast and the south face of the Cobequid Mountains is found a cover of surficial deposits (Swift and Borns, 1967a, 1967b). The basalt outcrops form upstanding masses reaching over 300 feet north-west of Cape Sharp and 200 feet in the case of Partridge Island. In both cases the slope down to the sea is steep.

The streams that flow from the Cobequid Mountains have steep -sided valleys in the mountains but they widen as they cut through the terrace between the mountains and the coast. In their lower parts, where the valleys are flat-floored, there is marsh land on either side of the river, for example Fox River, Ramshead River, and Diligent River.

Another notable point is that several of the rivers have long spits across their mouths, for example Greville River (Fig. 25), Fox River and Diligent River. Another spit has grown across an embayment of the coast at Union Valley, and Partridge Island is connected to the mainland by a shingle bank.

Parrsboro to Economy Point

This is a variable section of coast due to the outcrop of bedrock of varying resistance to erosion. Rocks of Pennsylvanian age form part of the coast whereas in other parts outcrops of the Annapolis Formation and of Triassic basalts are found. The situation is complicated by the existence in some parts of thick deposits of surficial material.

Along much of the coast there is a narrow strip of Triassic rocks, either the basalt or the Annapolis Formation. Where the coast is developed on basalt, it is high and steep, for example south of Greenhill, at Wasson Bluff, at McKay Head, and on the south-west side of Gerrish Mountain. Five islands, a group of steep islands, form a westerly extension of Gerrish Mountain.

Elsewhere, the coast is not necessarily low-lying but the high cliffs like those cut into the basalt do not exist. Where the Triassic sandstones and conglomerates outcrop, the cliffs are being eroded rapidly by the sea, resulting in some steep cliffs (Fig. 26) and peculiar rock formations, for example south of Lower Economy (Fig. 27). At high -tide the water level is above the base of the cliffs, but although they are undergoing rapid erosion, there are few boulders on the beach. This is presumably a result of the friable nature of the sandstone and its rapid break-up into its constituent particles. Where the cliffs are composed of surficial deposits, they usually have an irregular, slumped appearance.

This area is close to one of the extremities of the Bay of Fundy and the tidal range is extreme. Thus, there are extensive areas of tidal flats, for example at inlets at Five Islands and Lower Five Islands. Also, in the bay between Gerrish Mountain and Economy Point there is a wide expanse of tidal mud-flats and behind them, in a few localities, grass-covered marshes strewn with logs that have been floated in at high-tide.

Cobequid Bay

Cobequid Bay is taken to extend from Economy Point to Truro (north shore) and from Truro to Burntcoat Head (south shore). The Bay narrows to the east and it is in this area that the tidal fluctuation is greatest.

The enormous tidal range causes a complete transformation of the Bay from low-tide to high-tide. At low-tide it is occupied only by a narrow channel of water and is scarcely recognizable as an arm of the sea (Fig. 28). At high-tide, however, the Bay is occupied by red, muddy water. The two main rivers that flow into Cobequid Bay are the Salmon River from the east and the Shubenacadie River from the south; both are tidal some distance inland and the incoming tide causes a bore on the Salmon River.

Most of the coast of Cobequid Bay is cut into rocks of the Annapolis Formation, mainly soft red sandstone and conglomerate. The sandstone is easily eroded and provides the large quantities of mud that settle on the bottom of the Bay. The rapidity of erosion of the sandstone can easily be seen. From the top of the cliffs when the tide is in and the sea is lapping the base, the dark red coloration of the water near the cliff is apparent. The rapidity of erosion results in steep cliffs. Because the sandstone is soft, it breaks up rapidly into particles of sand and mud size so that few boulders are left at the base of the cliffs. Since the sea also picks out minor differences in resistance, arches and caves have been formed; also between Lower Truro Station and Clifton the cliff line truncates a number of rounded hills, referred to by Stevenson (1958) as "false drumlins."

Most of the rivers flowing into Cobequid Bay from the north have a common characteristic — near their mouths the valleys widen into marshy plains, in some cases dyked and in others not. Marshes occur at the mouth of Economy River, Bass River, Portapique River, Great Village River, Folly River and Debert River. Some rivers that enter from the south are similar, for example East Noel River. In contrast, the Shubenacadie River has steep cliffs on either side of it almost to its mouth, this despite the fact that there is very little water in it at low-tide (Fig. 29). The cliffs are cut into rocks of the Windsor and Horton Groups (Mississippian) which are more resistant than the sandstones of the Annapolis Formation. Mississippian rocks also outcrop between Selmah and Lower Selmah but do not produce any marked change in the nature of the coast.

Burntcoat Head to Cambridge Cove

Along most of this section of coast rocks of the Annapolis Formation are exposed and there is little difference between the coast and that of Cobequid Bay except that the tidal flats are not as extensive. The Annapolis Formation has been eroded to form some steep cliffs and differential erosion has led to the formation of caves (Fig. 30), stacks and conglomerate ledges some distance from the cliffs (Fig. 31). The rate of retreat of cliffs formed of rocks of the Annapolis Formation is rapid; small hanging valleys are common indicating that downward erosion by the streams has been unable to keep pace with the retreat of the cliffs.

The Annapolis Formation occupies only a narrow strip along the coast and where streams have produced inlets rocks of the Horton Group are exposed. In some places, for example Clement Cove, the marked unconformity between the steeply dipping Horton Group and the more nearly horizontal Annapolis Formation is clearly seen. At the mouth of the Tennycape River, rocks of the Horton Group have extremely high dips but have been planed off to form a flat surface.

Cambridge Cove to Avonport Station

Part of the south shore of Minas Basin, this area is developed almost entirely on rocks of the Horton and Windsor Groups. Although the tidal fluctuation is not as great as in Cobequid Bay, it is still large, and tidal mud-flats with marsh (dyked and non-dyked) are common (Swift, McMullen and Lyall, 1967; Swift and McMullen, 1968). Several large rivers flow to Minas Basin, the main ones being the Avon, Kennetcook, St. Croix, and Cogmagun. Their character is much the same as that of the tidal rivers described earlier.

The nature of the land/water contact varies; for example, whereas north of the mouth of the Kennetcook River the land rises gently inland (Fig. 32), at White Head near Cheverie and at Aberdeen Beach the shore is backed by prominent cliffs cut in white gypsiferous rock. Steep cliffs also occur at Horton Bluff where they are cut into bedrock, but to the north-west, at Avonport Station, the cliffs are composed of surficial deposits. Since the large boulders on the beach clearly do not come from the cliffs and since some of them are of amygdaloidal basalt, they were probably derived from North Mountain. Some of the boulders, found well below high-tide level and some distance from the cliffs, are isolated and surrounded by mud; thus it may be that they were floated in by ice, an explanation suggested by Dionne (1967) for similar boulders in the St. Lawrence Estuary.

Avonport Station to Borden Brook

Almost the entire coast is developed on soft Wolfville Sandstone and Blomidon Shale of the Annapolis Formation. Several rivers flow to Minas Basin the most southerly being the Gaspereau which is atypical because, for most of its length, it flows across rocks which are much more resistant than those of the Annapolis Formation. It, therefore, has a steep-sided valley and only near its mouth, where it flows across the Wolfville Sandstone, does the valley open out. The other rivers (the Cornwallis, the Canard, Habitant Creek, and Pereau Creek) flow across the Wolfville Sandstone or Blomidon Shale and, although their valleys are clearly-defined, they are not so steep-sided as that of the Gaspereau. All are tidal to a point well inland, the Cornwallis being dyked as far inland as Kentville. A re-entrant in the coastline at the mouth of each river is occupied by tidal salt-marshes, which are separated by headlands formed of the Wolfville Sandstone; Starr Point, Porter Point, and Longspell Point. Although the sandstone is easily eroded to give steep cliffs, it is uncommon to find sandstone boulders on the beach: basalt boulders, presumably derived from North Mountain, are more common.

Paddy Island is located close to the junction of the Wolfville Sandstone and Blomidon Shale. At high-tide an island, it is easily reached at low-tide by walking across the exposed sand-flats. Differential erosion by the sea has produced some unusual forms (Fig. 33).

Borden Brook to Cape Split

The coast is developed entirely on rocks of Triassic age, but the two groups involved are of markedly different lithology. The older Blomidon Shale consists of red shales and argillaceous sandstones which are easily eroded. Above it lies the resistant North Mountain Basalt (Fig. 34). The basalts are responsible for the upstanding land-mass known as North Mountain that extends from Cape Split to Digby. North Mountain has steep, south-east-facing slopes and a gentle north-west slope except at its extreme northern end where the rocks have been folded into a syncline so that the steep slope swings round to face north-east at Cape Blomidon and north-north-east at Cape Split. North of Borden Brook, the Blomidon Shales are capped by basalt and the cliffs become high and steep . Where the cliffs are cut into shale and sandstone, masses of rock falling from the cliffs remain on the wave-cut platform for a short time but there is a noticeable scarcity of shale and sandstone boulders. Northward, near Cape Blomidon, basalt boulders cover most of the wave-cut platform. Further north-west, the stage is reached where basalt is exposed in the cliffs, and at Cape Split, North Mountain tapers to a point and then to a series of stacks (Fig. 35)

Cape Split to Victoria Beach

The basalts of North Mountain slope gently to the Bay of Fundy and, with a few exceptions, the coast is developed on basalt. One of the most obvious results is that south of Scots Bay the coastline is remarkably straight. In places, especially in the south, the basalt dips gently beneath the sea; there is no real cliff (Fig. 4), and the slope of the lava flows is easily recognizable from the topography. Elsewhere, cliffs are developed and it is worth noting that sometimes there are two levels to the wave-cut platform, corresponding to individual lava flows. This point is mentioned here in order to draw attention to the possible dangers of jumping to quick conclusions about changes of sea-level on the basis of purely topographic evidence. If such an arrangement were found above present sea-level, it would be tempting to talk of a raised cliff. In parts the cliffs are steep, backing extensive wave-cut platforms covered by large boulders. Some of the boulders are extremely large; for example, near Ogilvie they reach 10 feet in diameter. One of the main characteristics of the cliffs is that they are under-cut at the base, the notch being curved in profile (Fig. 36).

Only at Scots Bay and between Margaretsville and Port George is the coast not formed on basalt. At Scots Bay some younger Triassic sedimentary rocks - the Scots Bay Formation (Fig. 8) - have been deposited on top of the basalt (Ells, 1893-1894; Haycock, 1899-1900; Powers, 1916; Klein, 1962). Scots Bay is also one of the few locations where marsh occurs. South of Margaretsville a section of the coast is cut into surficial deposits (Hickox 1958, 1962a, 1962b, 1966).

Dozens of streams flow to the coast down the north-west face of North Mountain. Almost without exception they are fast-flowing and have steep-sided valleys. In a few cases where coastal erosion has been rapid, downward cutting by the streams has not kept pace with the result that they reach the coast and fall over the cliffs from a considerable height.

Annapolis Basin and River

North Mountain is a continuous area of highland from Cape Split to Digby Gut where, cutting across the Mountain, there is a deep gap whose direction is fault-determined. The water occupying the gap has a depth of 309 feet at lowest normal tides (Canadian Hydrographic Chart 4396). Inland from the gap, a wide basin, Annapolis Basin, is developed on the soft rocks of the Annapolis Formation. The Annapolis River flows from the north-east gradually widening until it merges with Annapolis Basin. The river was tidal to a point well above Bridgetown, as is evidenced by the dykes along the river, but it is tidal no longer as a control dam has been built at Granville Ferry. Both above and below the dam there are extensive areas of reclaimed marsh, for example Pré Rond Marsh.

Digby Neck, Long Island, and Brier Island

A continuation of North Mountain, these are composed almost entirely of North Mountain Basalt, but there are a number of gaps through the basalt outcrop, one of them being Digby Gut which has already been mentioned. Two more reach below sea-level, namely Petit Passage between Digby Neck and Long Island, and Grand Passage between Long Island and Brier Island. The three gaps are fault-determined and it is noticeable that in each case there is an offset in the coastline (Fig. 4). In addition, there are other gaps which do not reach below sea-level, for example at Gulliver Cove, Centreville, and Sandy Cove. Other gaps, too small to mention by name, can be recognized on the 1:50,000 topographic maps.

The basalt flows are of varying resistance; one flow of lesser resistance is exposed along the centre of the peninsula and of the islands resulting in a valley which runs parallel to the coastline on either side (Fig. 37). Midway Lake, south of Centreville, occupies part of this lowland and a deep inlet of the sea, south of Freeport on Long Island, is developed along the same line, as is Pond Cove at the southern tip of Brier Island. Where cliffs have been cut into the basalt, they are steep and in a few places show the characteristic columnar jointing of basalt, for example at East Ferry on Digby Neck. Sandstones of the Annapolis Formation are exposed near Rossway and give rise to low cliffs.

Coastal deposition is not common but East Sandy Cove is largely an area of mud-flats at low-tide and there is a well-developed bay bar at Pond Cove.

St. Marys Bay (excluding Digby Neck and the islands)

At the northern end of the Bay, rocks of the Annapolis Formation outcrop and give rise to a low-lying coast, whereas most of the east coast is developed on much older rocks of the Goldenville Formation (Ordovician). Although this is a low-lying, irregular coast, it would be much more irregular but for coastal deposition which has formed spits and bars across some inlets, for example at Ticken Cove and south of Saulnierville. Other large spits occur at Gilbert Point, Weymouth Harbour, and south of Beliveau Cove.

At the southern end of the Bay irregular cliffs and a wave-cut platform have been cut into the Ordovician rocks. Johnson (1925, p. 161) stated that this area shows "what appears to be the 'crenulate shoreline' typical of a very youthful shoreline of submergence" and that "sea cliffs are normally minor scarps cut but a short distance into much higher hill-slopes of the partially submerged landmass."

Summary of the Characteristics of the Fundy Coast

In summary, it has been shown that the Fundy coast exhibits a great variety of forms due to variations in geology (lithology and structure), exposure, and tidal range. The rocks range from extremely resistant igneous rocks, as between Deadman Head and Lepreau Harbour, on which the sea has made little impact at present sea-level, to easily-eroded sandstones and shales which are being rapidly removed, for example around Cobequid Bay and much of Minas Basin. Some parts such as along the whole of North Mountain and Digby Neck, are parallel to the structural trend, whereas others are at right angles to the structure, for example at Salisbury Bay. Some parts of the coast are sheltered, like Passamaquoddy Bay; the other extreme is represented by places like Cape Enragé and Cape Chignecto which are open to the maximum fetch along the south-west/north-east axis of the Bay. At the mouth of the Bay where tidal range is low by Fundy standards, salt -marsh accumulation and tidal mud-flats and sand-flats are of limited distribution, but at the extremities - Shepody Bay, Cumberland Basin and Cobequid Bay - there are many square miles of marshes, both reclaimed and tidal, and of tidal sand- and mud-flats. The marshes and mud-flats are one of the unique features of the coast. As will be shown later, they are useful when trying to unravel the geomorphological history of the Bay, although the reclamation of some and not of others is a complicating factor.

Three other characteristics of the coast are worth noting at this point:

(1) On the New Brunswick side are several small outliers of Triassic rock which, although limited in area, are important in determining the erosional and depositional history of the Bay.

(2) All the rivers entering the Bay from the St. Croix to the St. John (inclusive) have falls at their mouths. Although some other rivers have falls too, the phenomenon is nowhere else as uniform as along the south-west New Brunswick coast.

(3) Coastal depositional features - spits, bars, and tombolos - are common along the New Brunswick coast and along the Nova Scotia coast of Chignecto Bay and the north shore of Minas Channel, Minas Basin and Cobequid Bay. In contrast, they are virtually absent along the Nova Scotia coast from Truro to Brier Island the exception being in St. Marys Bay.

Submarine Topography and Sub-Bottom Deposits

Any study of the Bay of Fundy must be regarded as a study of part of the continental shelf. Therefore, in addition to a description of the submarine topography of the Bay itself, a brief description of the adjacent parts of the continental shelf is included here.

Before the advent of modern depth-finding techniques, the general nature of this part of the continental shelf was known. Johnson (1925, pp. 264-265) wrote:

> Between the rocky New England coast where the dissected peneplane of the crystalline upland slopes down into the sea, and the remarkable series of "banks" lying approximately 150 miles off that coast, there is a body of water in places over 1000 feet deep, which is called the Gulf of Maine. The great basin-like depression which holds this water body is continued on the northeast by the Bay of Fundy lowland until the floor of the latter rises slightly above sealevel in the narrow Isthmus of Chignecto. To the southwest it is cut off where its submerged southeastern rim, formed by the Banks, converges to meet the rocky northwestern rim at the base of Cape Cod. The only effective outlet from the basin is a channel 800 or 900 feet deep between Georges Bank and Brown Bank.

A map of submarine topography between New Jersey and Nova Scotia compiled by Uchupi (1964a) shows that the continental shelf ends abruptly seaward in a steep slope, the continental slope, which is dissected by a number of steep-sided submarine canyons and that Georges Bank and Browns Bank, which separate the Gulf of Maine from the continental slope, are areas of shallow water, in parts less than 300 feet deep.

A channel, separating Georges Bank and Browns Bank and referred to as Northeast Channel, leads from Georges Basin, a narrow depression within the Gulf of Maine, to the top of the continental slope (Fig. 38). Another less well-defined channel, Great South Channel, leads southward from the Gulf of Maine towards Hydrographer Canyon which is incised into the continental slope. The submarine topography of the Gulf of Maine is by no means uniform, for it consists of a series of basins separated by banks and ridges (Uchupi 1965, 1966a).

Viewed in general terms the Bay of Fundy, which is an arm of the Gulf of Maine, decreases both in width and in depth towards the north-east. This shoaling towards the north-east is shown well on Canadian Hydrographic Service Charts no. 4010 and no. 4011. Despite the small scale of these charts (1:200,000 and 1:300,000 respectively) they show that, although there is an overall shoaling to the north-east, there are some well-marked basins and ridges in the Bay.

The existence of several deep-water channels within the Bay has long been recognized and there has been considerable discussion of possible modes of origin. Matthew (1879, pp. 18EE-20EE) referred to eight such channels and attributed them to erosion by tidal currents. Much later, Shepard (1930, 1942) cited depressions in the floor of the Bay as evidence in favour of a glacial origin. In a much more recent study, Swift and Lyall (1968b) referred to bedrock troughs in Chignecto Bay, Minas Passage (between Cape Split and Cape Sharp), and at the mouth of the Bay of Fundy which they concluded could not be attributed to modern tidal scour. Sub -bottom profiles also revealed "a buried bedrock river channel ... which runs from Chignecto Bay past Isle [sic] Haute to the south shore." (Swift and Lyall 1968b, p. 342).

In addition to deep-water channels and basins several prominent ridges are referred to in the literature (Bailey, 1904, 1910, 1919; Johnson 1925). The early references to ridges were based on limited surveys but recently Swift and Lyall (1967, 1968a, 1968b) obtained cross profiles of the Bay which confirm their existence. One section, passing through Ile Haute, shows that the island is part of a basalt ridge that extends westward to Quaco Ledge, 10 miles east-south-east of Quaco Head. Another basalt ridge further north does not project above sea-level. Two sections, drawn nearer the mouth of the Bay, show no evidence of the basalt ridges.

For many years researchers have speculated about the lithology of the floor of the Gulf of Maine and the Bay of Fundy. Johnson (1925, Fig. 144) showed Triassic rocks extending out from the Bay of Fundy into the Gulf of Maine. Not until the development of modern seismic and sonic techniques, however, could any definitive statement be made. Drake, Worzel, and Beckman (1954) showed cross profiles of the Gulf of Maine,obtained by using seismic refraction measurements, which confirmed the existence of Triassic sediments beneath the Gulf. Likewise, Tagg and Uchupi (1966, Fig. 28) showed that Triassic sediments extend southward from the Bay of Fundy into the north -eastern part of the Gulf of Maine. Swift and Lyall (1968a, p. 641) stated that "except for a zone on the north shore about 10 km wide, the Bay of Fundy is floored by gently dipping to horizontal Triassic sedimentary rocks."

Some attention has also been given to the nature of the surficial deposits. Hachey (1952) distinguished three categories of surface material within the Bay of Fundy: a) mud, b) sand and stones, and c) rocks and ledges. Forgeron (1962, p. 114) summarized the nature of the Bay of Fundy floor as follows: "The greater part of the Bay of Fundy is underlain by gravel and rocky bottom with fine sediments occurring mainly on the Saint John delta, off the mouth of Chignecto Bay, and in the eastern central Bay of Fundy." The three cross profiles drawn by Swift and Lyall (1967, 1968a, 1968b) show a thick layer of surficial deposits, the bulk of which is thought to have been deposited during the Pleistocene. They concluded that in many places the bottom materials are still overlain by outwash. Some of the materials deposited during the Pleistocene have been redistributed during post-glacial time, an example occurring in Minas Channel where a sub-tidal sand body has been "localized by the interaction of the bottom topography with the tides "(Swift, Cok, and Lyall, 1966, p. 175)

Although they cannot correctly be termed submarine, in that they are not constantly covered by water, the tidal deposits of the Bay of Fundy should be mentioned because they are extensive and a distinctive aspect of the Bay of Fundy coast with careful interpretation they can be a useful indication of changes in sea-level. Given a slight rise in sea-level, extensive areas of tidal deposits would be permanently covered by the sea, but with a slight drop in sea-level they would soon be transformed into well-drained land.

Many of the areas where tidal deposits occur have already been described in an earlier section of this thesis, but a few areas deserve individual mention at this time. Six areas display well the characteristics of the tidal deposits of the Bay of Fundy:

(1) the Musquash region south of Saint John,

(2) the head of Shepody Bay,

(3) the head of Cumberland Basin,

(4) Cobequid Bay,

(5) the south shore of Minas Basin,

(6) the shore of Annapolis Basin.

In all cases the natural situation has been partly disturbed by reclamation of the higher parts of the marshes.

Tides

The Bay of Fundy has the greatest known tidal range in the world. This phenomenon has attracted much interest, not only from natural scientists but also from engineers who have been concerned with the possibility of harnessing the tides to produce power. So far their investigations have taken the form of feasibility studies (Smith, 1958).

Hamilton (1867) wrote the first detailed account of the tides, dealing at some length with the existence of tidal currents within the Bay and with the phenomenon of "the bore". On the basis of reports from people living on the shores of Minas Basin, he stated that a tidal range of 75 feet was not uncommon and that in the past this was the usual tidal range at spring tides. Hind (1875) described the Fundy tides, his main concern being their effect on the feasibility of building a canal across the Chignecto Isthmus. In his account he reported the tidal range at spring tides as being 48 feet at the mouth of the Avon River and 65 to 70 feet at the mouth of the Shubenacadie further east and nearer the head of Cobequid Bay. Chalmers (1895) mentioned the existence of a bore on the Petitcodiac River and another which is less well-known on the Maccan River. These, he said, are due to the narrowing of the upper part of the Bay. Goldthwait (1924) also included a section on the tides of the Bay of Fundy and drew a map showing the tidal ranges in the Bay (Fig. 39). Two years later, Kindle (1926) dealt with certain results of the large tidal range: 1) the effervescence or bubbling of water as the tide covers previously exposed sand areas; 2) the formation of sand waves and "split bars"; and 3) the formation of tidal waves or bores.

It can be seen from Goldthwait's map that the tidal range increases from the mouth of the bay towards Chignecto Bay and Minas Basin. It is in Minas Basin that the greatest tidal ranges occur. The *Atlantic Coast Tide and Current Tables* (1965, p. 3) state that:

> It is off Burntcoat Head, in the upper part of Minas Basin, that the largest tides in the world have been recorded. On July 16 and 17, 1916 tide ranges of over 53 feet were measured. The average range of the lunar semidiurnal tide at this point is about 40 feet, while that of the average solar tide is only about 6 feet.

It is also noticeable that the tidal range is greater along the south-east coast of the main part of the bay and the south coast of Minas Basin than on the New Brunswick side.

The large tidal fluctuation is responsible for two of the most distinctive features of the rivers which flow into the Bay of Fundy. At the head of the bay, three rivers are affected by tidal bores: the Petitcodiac, the Maccan, and the Salmon. On the Petitcodiac the bore reaches Moncton some three and a half hours before high-water, and occasionally at times of high-tide it is reported to be as much as 4 feet in height. However, normally it is more in the order of 6 inches to 1 foot high.

At the Reversing Falls at the mouth of the Saint John River, the level of water in Saint John Harbour at low-tide is below the water level in the river above the gorge so that water rushes through the gorge at great speed. However, at high-tide the water level in Saint John Harbour is higher than that in the river. Consequently, water rushes landward. There is a period between high-tide and low-tide when the level above and below the gorge are about the same and at that time there is no appreciable movement in either direction.

Abnormal tides occur on rare occasions. Three of the best documented occurrences are: the storm of November 3, 1759 (Hind, 1875); the "Saxby Storm" of October 5, 1869 (Hind,1875; Ganong,1911); and a gale on November 9, 1900 (Stead, 1903). The possible significance of these events is discussed later in Chapter 5.

In this section on tides no attempt has been made to explain the large tidal range of the Bay of Fundy. This aspect will be dealt with later when discussing the factors that could have caused the tidal situation to be different in times past.

CHAPTER 3

PRE- QUATERNARY DEVELOPMENT

A complete study of the Bay of Fundy trough would involve investigations extending back to the earliest geological time. There are several references in the literature to the great age of a geosyncline or trough existing in the general area of the present bay (Bailey, 1897; Daly, 1901; Schuchert and Dunbar, 1934; Cameron, 1956a). The general consensus of opinion in the literature is that rocks of Cambrian to Carboniferous age were deposited in the geosyncline or trough.

After the Carboniferous the next youngest rocks are of Triassic age. These are exposed extensively along the Nova Scotia coast of the Bay but are found only in isolated patches along the New Brunswick side (Fig. 6). There follows a great gap in the geological column and, apart from some deposits in the Shubenacadie Valley that contain Early Cretaceous flora (Klein 1960), the next youngest rocks exposed on land are of Pleistocene age. Some doubt exists about the geological age of the "Bridgewater Conglomerate" of southern Nova Scotia, but Grant (1963) argued conclusively that it is of Pleistocene age.

Growing evidence indicates that rocks of Cretaceous and/or Tertiary age occur below sea-level off the south-east coast of Nova Scotia (Officer and Ewing, 1954; Marlowe, 1965; King, 1969) and beneath Georges Basin (Uchupi, 1966a) and Northeast Channel (Ballard and Sorensen, 1968; Uchupi, 1966b)

Conditions under which Triassic Rocks Were Formed

As this study really starts at the beginning of the Triassic, an attempt will be made to trace the geomorphological history of the Bay of Fundy from that time onwards, although, of necessity, studies of the Pleistocene will be much more detailed than those of earlier times.

It is important to understand the conditions under which the Triassic rocks were formed because this gives a good indication of the palaeo -geography during Triassic times. Dawson (1848, 1868) believed that the Triassic rocks of the Maritime Provinces, with the exception of the basalts, were of marine origin and that a proto-Bay of Fundy was in existence when they were deposited. Powers (1916) made a detailed study of the structural and stratigraphical relations of the various Triassic horizons. He recognized the continental origin of the Fundy Group red beds and inferred that deposition took place in an intermontane fluvial environment subjected to an arid and semi-arid climate; he also suggested a lacustrine origin for the Scots Bay Formation. Recently Klein (1963a) conducted a painstaking study of the dip directions of cross-stratification in the Fundy Group and on the basis of his results made the following comments about the conditions of formation:

> The broad pattern of sedimentation in the northeast-trending fault trough in the Canadian Maritime Provinces was one of basin filling from both the north and northwest and south and southeast during the Triassic period. This pattern persisted during the deposition of the entire Triassic sequence in New Brunswick. In Nova Scotia, this pattern of stream drainage flowing from the north and south of the basin is characteristic of both the Wolfville and Blomidon Formations. The volcanism extruding the McKay Head basalt dammed the local drainage to form the lake in which the Blomidon was deposited but did not otherwise disrupt the drainage pattern. When the North Mountain Basalt was extruded, however, the drainage flow pattern from the north and south of the basin was disrupted completely and only a local lake deposit was formed (Scots Bay Formation). Its source of sediment is not known (Klein 1963a, pp. 803-804).

Klein, then, in common with Dawson, recognized the existence of a trough in the general area of the present Bay of Fundy during Triassic times although the size and shape were somewhat different from those which exist at present.

Structure

Knowledge of the existence of major structures in the Maritime Provinces is of importance in unravelling the geomorphological history of the area. Hobbs (1904) attempted to distinguish "lineaments" or structural lines determining the present configuration of the "Atlantic Border Region". One such line was supposed to delimit the south-eastern coast of the Bay of Fundy. Another runs along the western coastline of Nova Scotia through Petit Passage and is continued northward by the zigzagging course of the Saint John River. Another is a line followed by the St. Croix River.

Since Hobbs' work appears to have been based mainly on evidence derived from the alignment of topographic features, it is surprising that he made no mention of the area north-west of the Bay of Fundy. Here there are at least four depressions trending north-east/south-west which seem to be structurally determined. From north-west to south-east they are Grand Lake, Washademoak Lake, Belle Isle Bay - Long Reach, and Kennebecasis River and Bay (Fig. 6). The Bay of Fundy is sometimes regarded as a fifth in the series (Bailey, 1910).

Powers (1916) was, perhaps, the first person to show that the hook at the north end of North Mountain is due to a syncline which plunges to the west. He also recognized the existence of some important faults, among them: 1) a fault on Grand Manan Island which causes the downthrow of basalts to the west; 2) faults on the north and north-west sides of Triassic outcrops at Quaco Bay, Martin Head, and Waterside; 3) the Cobequid fault, an east-west trending fault with a displacement of 2000 feet - 3000 feet; and 4) cross faults at Digby Gut, Gulliver Cove, Petit Passage,and Grand Passage.

Goldthwait (1924) also attributed the gaps in North Mountain and in Digby Neck to faulting, the lines of weakness subsequently being occupied by rivers. He suggested that the North Mountain syncline stretches east to Truro, affecting the red sandstones that underlie the basalts of North Mountain.

Johnson (1925) distinguished a "Fundian Fault" which he thought determined the alignment of the north-west Fundy coast. He explained the steepness of parts of the north-west shore of the Bay of Fundy by referring to the fault, but Shepard (1930) ascribed the steepness to glaciation. The latter view was opposed by Koons (1941a, 1941b), a disciple of Johnson, who cited further "evidence" for the fault theory. Later articles by Shepard (1942) and Koons (1942) did little to resolve the argument.

The work of Cameron (1949, 1956a)contributed to the understanding of the tectonics of the Maritime area which he saw as lying between the Canadian Shield on the north and west "the general boundary being a thrust fault of unknown but probably large displacement" (1956a, p. 45), and, to the south-east, the edge of the continental shelf where the boundary "is inferred to be a thrust fault similar to that on the north and west" (1956 a, p. 47). The Appalachian geosyncline extended between the two boundaries and in it deposition and folding took place, the folded rocks forming the source of sediment for subsequent deposition. He showed a number of faults north-west of the Bay of Fundy trending at an angle not quite parallel to the Fundy coast and also two fault lines running along the Bay of Fundy (1956a, Fig. 2).

Swift and Lyall (1968a), by investigating the submarine geology of the Bay of Fundy, threw more light on the structure of the area, particularly on the "Fundian Fault". They traced the Fundy syncline out of Scots Bay into Minas Basin and as far south as the mouth of the Bay of Fundy. They also distinguished a series of discontinuous faults forming a zone along the north-west side of the Bay. This zone resembles the "Fundian Fault" of Johnson although it lies further south-east. They concluded that:

> The bay is underlain mainly by Triassic sedimentary rocks. Fundy's shoreline is nearly coincident with the Acadian Triassic basin. An offset of some 10 km between the topographic and structural basins leaves Triassic rocks exposed on land to the southeast and has submerged pre-Triassic rocks on the northwest. The subaerial southeast margin of the structural basin consists mainly of an unconformable contact between Triassic and pre-Triassic rocks while the submarine northwest margin is defined by a series of faults. The basin is, therefore, a half-graben. However, the profiles suggest that the great central syncline is a more significant structure and show that the axis of the syncline is displaced to the southeast rather than to the northwest of the basin's topographic midline. The syncline is, therefore, not closely controlled by the fault zone and the Acadian Triassic basin is more truly a structural basin than it is a fault trough (Swift and Lyall 1968a, p. 645).

Peneplains and Erosion Surfaces

Historical Outline

Daly (1901) was the first to interpret the physiography of Acadia in terms of peneplains. The oldest of these, represented in part by the Atlantic Uplands, the Cobequid Mountains, the Southern New Brunswick

Highlands (St. Croix Highlands and Caledonian Highlands) (Fig. 5) and the surface of North Mountain, Digby Neck,and Long Island, he termed the "Cretaceous Peneplain". He thought that the peneplain had been tilted in a direction about south-30 degrees-east with a secondary warp transverse to this, and that if the surface of the Atlantic Uplands was extended in imagination to the Southern New Brunswick Highlands, it would be tangent to the summit of North Mountain. The completion of the formation of the peneplain was given as the close of the Cretaceous Period, a date which was obtained by comparison of the situation in the Maritime Provinces with that in "better known regions" to the south. Using the same line of reasoning, he gave the age of tilting and warping of the peneplain as late Cretaceous or early Tertiary. As a result of the tilting and warping of the peneplained surface a new cycle of erosion was initiated and lowlands were formed on the softer rocks of the region, the harder rocks being left as upstanding masses. Daly believed that the lowlands belonged to "one great plain of denudation dating from the end of the Tertiary cycle" (Daly 1901, p. 95). This plain he termed a secondary peneplain, or "Tertiary Peneplain". Like the Cretaceous peneplain it was uplifted, probably with differential movement during the late Tertiary.

Assuming for the moment that Daly's correlation of the remnants of the Cretaceous peneplain is correct, the main weakness of his work is the false impression given by his use of the word "peneplain" for the lowlands eroded on the Carboniferous and Triassic rocks of the area. Davis (1895, p. 497) in his classic essay on the geographical cycle wrote: "An almost featureless plain (a peneplain) showing little sympathy with structure, and controlled only by a close approach at baselevel, must characterize the penultimate stage of the uninterrupted cycle." It can hardly be argued that a lowland developed on the Carboniferous and Triassic rocks of the Maritime Provinces, interrupted by North Mountain, the Cobequids, and the Southern New Brunswick Highlands, fits this description. The existence of these highlands reflects an adjustment to structure which is scarcely compatible with the Davisian concept of the peneplain.

Once established in the literature the peneplain concept was hard to remove. Ganong(1902a) referred to a Cretaceous peneplain and a Tertiary peneplain; Bailey (1910, p. 59) to "two quite distinct and successive periods of peneplanation, one completed in the Cretaceous and one in the Tertiary."; and Powers (1915) believed that two peneplains had been developed and uplifted between the end of the Triassic and the late Tertiary. In his monograph, *Physiography of Nova Scotia*, Goldthwait (1924) referred to the "Atlantic Upland" rather than to a Cretaceous peneplain, but apart from this and some other minor differences in terminology, he described the sequence of events suggested by Daly twenty-three years earlier. Johnson (1925) in *The New England-Acadian Shoreline* accepted the first part of the sequence; that is, the formation of a plain across all rocks irrespective of their resistance, but thought that it might be of later date than the Cretaceous. For this reason he referred to it as the "Acadian Peneplane", thus eliminating any implication as to its date of formation.

The concept of a peneplain (or peneplains) in the Maritime Provinces was perpetuated by Hayes and Howell (1937); MacKenzie (1940); and by Koons (1941a, 1941b) who referred to a "Summit Peneplane" which bevels the North Mountain-Digby Neck-Long Island-Brier Island ridge. Although the word "peneplain (peneplane)" has become less common in the literature on the

Maritime Provinces since the early 1940's, it has not been eliminated completely (Hudgins, 1960; Cumming, 1967).

Upland surfaces and uplands are still recognized. Bird and Hare (1956) questioned the correlation by Goldthwait of the erosion surface of the Atlantic Uplands of Nova Scotia with the surfaces of the Cobequid Mountains and of the Caledonian Highlands. They suggested that the surface truncating the shales and quartzites of the Atlantic Uplands of Nova Scotia "was independently tilted, possibly in the remote geological past, buried and then exhumed from the later Paleozoic rocks " (Bird and Hare 1956, p. 42). Bird (1964) thought that the dominant uplands below 2200 feet in the Maritime Provinces represent a warped pre-Tertiary surface; that the initial drainage was by south and south-easterly flowing streams; and that this pattern was disrupted by the development of lowlands on weak Carboniferous sediments. He believed that exhumed surfaces are widely distributed, and that "the largest that is at least in part of Sub-Triassic origin forms a rugged, tilted zone, 40-80 km. wide along the Atlantic coast of Nova Scotia " (Bird 1964, p. 119).

Examination of Map Evidence

Much of the literature about peneplains in the Maritimes is based on now-abandoned assumptions and on data derived from very limited observations; for example, Daly (1901) based his conclusions on observations made almost entirely from railway tours. At that time, of course, topographic map coverage was limited and restricted to small scale maps. Consequently, the peneplains referred to by Daly, and subsequently accepted almost without question by many others, need some detailed study.

The area is now covered by 1:50,000 topographic maps but there has been little detailed study aimed at determining whether the peneplains (erosion surfaces) really exist or whether they are figments of a fertile imagination. Consequently, the writer carried out three items of map analysis aimed at clarifying the situation:

(1) 5 cross profiles of southern New Brunswick and Nova Scotia were drawn using 1:50,000 topographic maps;

(2) cross profiles of North Mountain were drawn from 1:50,000 topographic maps;

(3) height analysis of southern New Brunswick and mainland Nova Scotia was carried out using the 1:50,000 topographic maps.

Cross profiles of southern New Brunswick and Nova Scotia

The five cross profiles were drawn in a north-west/south-east direction at right angles to the structural trend (Fig. 40). It was reasoned that if there is any correlation between the summits of the highlands of southern New Brunswick and mainland Nova Scotia this might show up on the sections. The horizontal scale used was 1:50,000 and the vertical scale, 1:6000, a vertical exaggeration of 8.3. The resulting sections are an unwieldy 20 feet in length but any reduction in the horizontal scale leads to gross vertical exaggeration. Therefore, the sections are not included as illustrations in this dissertation. However, bearing in mind the purpose for which they were constructed, the most significant features of the sections are noted as follows:

Section 1.

(1) There is a gentle slope to the south-east in New Brunswick, the height decreasing from 500 feet to 150 feet. If this slope is continued southward,

it passes far below the top of Dibgy Neck,which rises to 250 feet.

(2) On the Atlantic Uplands of Nova Scotia there is a plateau at 250 feet with occasional higher spots reaching 300 feet, and then a slope to the south-east from 250 feet to sea-level.

Section 2.

(1) On the New Brunswick side of the Bay the relief varies considerably with no obvious flat or gently inclined surfaces.

(2) North Mountain reaches a height of 500 feet.

(3) On the Atlantic Uplands there is, in the west, a plateau at 500 feet, then a sharp step down to 250-300 feet from which level there is an irregular drop to sea-level.

Section 3.

(1) In New Brunswick there is great variation of relief which can be explained by the fact that the lithology in this area is variable.

(2) North Mountain reaches a height of 800 feet; it is notable that South Mountain is at a lower altitude.

(3) The north-western part of the Atlantic Uplands is a plateau at approximately 600 feet with occasional higher points up to 750 feet and lower points down to 500 feet.

(4) South-east of this a gently-sloping surface decreases from 400 feet to 300 feet in a south-easterly direction.

(5) Further south-east there is a plateau at 200-250 feet and then a gentle drop to sea-level.

(6) Although there are interruptions, the Atlantic Uplands have an overall south-easterly slope.

Section 4.

(1) In New Brunswick the relief is very irregular except near the Fundy coast where there is a highland at 1000+ ft. with a tendency towards a south-east slope to the Bay of Fundy. If this slope were continued across the Bay of Fundy, it would pass below the summit of North Mountain.

(2) North Mountain reaches an altitude of 650 feet and has a definite cuesta form.

(3) South Mountain has an irregular rise up to 750 feet.

(4) A plateau occurs on South Mountain at 600-700 feet.

(5) In the south-east the drop to the coast is irregular.

(6) Along this section line the slope of the Atlantic Uplands would project well above North Mountain.

Section 5.

(1) The New Brunswick part of the cross profile is very irregular.

(2) The Cobequid Mountains show up as a plateau at 600 feet to 700 feet.

(3) Further south-east, at 500-600 feet, there is another plateau which has a definite slope to the north-west.

(4) In the south there is a gradual but irregular drop to the south-east.

Five cross profiles represent a very small sample of the south-west/north-east extent of the landforms across which they pass. With such a small sample local conditions may be more prominent than regional trends; for example, the section line might by chance run along a valley which has high land on either side. Perhaps a more valid method would have been to draw generalized contours and to compare summit levels. With this constraint in mind the following conclusions are drawn from inspection of the cross profiles:

(1) No evidence exists that the plateau surface of the Atlantic Uplands of

Nova Scotia, if projected to the north-west, is tangent to the top of North Mountain; nor is there any evidence that it correlates with an erosion surface in southern New Brunswick.

(2) The Atlantic Uplands do not have a uniform slope to the south-east. They consist, rather, of a series of plateaux and sloping sections often separated by steps, as suggested by Bird and Hare (1956).

(3) A uniform south-east slope is most marked in the south-eastern third of Nova Scotia.

Cross profiles of North Mountain

In addition to the sections of southern New Brunswick and Nova Scotia cross profiles of North Mountain (110 in all) were constructed, the object being to test the statement by Daly that North Mountain is a flat-topped, truncated ridge, the truncation being due to the Cretaceous peneplain. His idea persisted in the literature for at least 40 years (Koons 1941a, 1941b), but its validity was not tested as topographic maps were not available.

The profiles were drawn along the interfluves between the small rivers flowing from North Mountain north-west to the Bay of Fundy and south-east to the Annapolis-Cornwallis Valley and St. Marys Bay. This meant that they followed a general north-west/south-east direction at right angles to the structural trend except at the extreme northern tip of North Mountain.

Sections were drawn with a horizontal scale of 1:50,000 and a vertical scale of 1:6000, giving a vertical exaggeration of 8.3. In order to facilitate representation of these profiles and to attempt to get an overall view of the cross profile of North Mountain, projected profiles were used (Fig. 41). On each of the diagrams 10 profiles are projected, this being the maximum number than can be shown on one diagram without confusion. In each case the viewpoint is assumed to be from the south-west.

If North Mountain has been bevelled by a Cretaceous peneplain which was later tilted to the south-east, it is reasonable to expect evidence of flat surfaces sloping slightly to the south-east. On only 6 of the 110 profiles and 2 of the 11 projected profile diagrams (Fig. 41, nos. 4 and 5) is there any suggestion of this arrangement. Much the most common situation is a straightforward scarp and dip slope, the scarp facing south-east and the dip inclining to the north-west, although on some of the projected profile diagrams (Fig. 41, nos. 8, 10 and 11) there is evidence of a general flattening of the top of North Mountain.

The conclusion derived from inspection of the profiles is that some evidence exists for a truncation of North Mountain, but that there is little to suggest that this is anything to do with a south-eastward tilted Cretaceous peneplain.

Height analysis of southern New Brunswick and Nova Scotia

A simple height analysis of an area can indicate that within it there is a predominance of points at a certain height. Care must be taken not to read too much into the result unless detailed fieldwork has been done in the area. For example, on the basis of a predominance of points at say 300 feet it might be concluded that an old erosion level is represented when in effect it reflects the existence of an extensive outcrop of a resistant rock at this level. As the writer did little fieldwork in the interior of New Brunswick and Nova Scotia, any conclusion arrived at on the basis of this height analysis must be regarded as tentative.

The area covered by the height analysis is shown in Figure 42. Complete topographic map coverage of this area is available at a scale of 1:50,000 (165 maps in all), most of the maps having a contour interval of 50 feet, although in a few cases the interval is 25 feet.

The type of height analysis undertaken does not throw any light on the validity of Daly's tilted Cretaceous peneplain for the following reason. Assuming that it exists and that it is tilted downwards to the south-east, the remnants of it which might lie at 900 feet in southern New Brunswick will be equivalent genetically to surfaces at say 200 feet in south-eastern Nova Scotia. Therefore, even if a peneplain does exist there will not necessarily be any maximum of points at a specific height.

The method used was to construct on a sheet of tracing paper, large enough to cover one topographic sheet, a grid of horizontal and vertical lines spaced at one-inch intervals (which is equivalent to 4167 feet on the ground). The grid was placed over the map and the height at each intersection was recorded. The distance between the grid lines was arbitrarily chosen except that it gave a reasonably close network, considering the large area being covered, and yet it was not so close that the task of height determination was impossibly lengthy. As some of the maps had a 25-foot contour interval, the class interval used was 24 feet (0-24, 25-49, up to the highest points in the area, 1350-1374). Naturally the grid intersections did not frequently fall on a contour line and interpolation of heights was necessary, especially on the maps with a 50-foot contour interval.

The totals for each height range were obtained and the results plotted on a simple graph showing height range against frequency (Fig. 43). This was done for the whole area first and then individually for the upstanding physiographic divisions of this area (the Southern New Brunswick Highlands, the Cobequid Mountains and the Antigonish Mountains, North Mountain, and the Atlantic Uplands of Nova Scotia).

When a class interval of 24 feet is used, it is noticeable that above the class 75-99 the lower half of any 50-foot range tends to predominate; for example, heights in the range 100-124 are more common than those in the range 125-149; similarly, 150-174 is more common than 175-199. This trend is so marked that it is attributed to the fact that when an observer had to interpolate within a 50-foot range he usually estimated in the lower half of the range. This is evident despite the fact that many different observers carried out the height analysis.

In order to overcome this problem, a second graph was constructed with a class interval of 49 feet (Fig. 44). This graph shows that up to the class 100-149 there is an increase in frequency of points and then a steady decrease in frequency. There are three exceptions to this pattern: 1) class 200-249 is more common than 150-199; 2) class 450-499 is only slightly less common than 400-449; and 3) class 600-649 is only slightly less frequent than 550-599.

The area being subjected to analysis is one of variable relief with heights ranging from 0 to 1350 feet. It has a long coastline and is drained by many rivers which vary greatly in valley size. Under these circumstances it is predictable that there will be a maximum frequency of heights in the lower ranges and then a steady decrease with height, unless the arrangement is upset by the existence of extensive erosion surfaces or structurally-determined terraces. Although it is predictable that the maximum frequency of points will be in the lower height ranges, it is not likely to be the class 0-49 because these heights will occur only close to the coast or inland in the valleys of major rivers. The maximum at 100-149 shown in Figure 44 is interpreted as being composed of points close to, but not on, the coast; points in the valley bottoms of major streams some distance inland; and points on valley sides. The fact that the steady decrease in frequency with height is interrupted at 200-249, 450-499, and 600-649 is thought to be of significance. These may represent old erosion levels, but for reasons outlined earlier the writer is not prepared to be any more specific.

In order to carry the analysis further and to see whether there are predominant height ranges in the upstanding physiographic regions of the area under study, four separate graphs were constructed for the Southern New Brunswick Highlands, the Cobequid Mountains and the Antigonish Mountains, North Mountain, and the Atlantic Uplands of Nova Scotia (Fig. 45). The method used was not entirely satisfactory, so the conclusions that can be drawn from the results are limited. The boundaries of the physiographic regions in question were not marked on the topographic maps before analysis started. Thus, when the number of points for the various heights within a given physiographic region were totalled, all the heights obtained from any map which included part of that region had to be included. This meant that heights for land outside the physiographic region were included. This problem was greatest in the case of the narrow physiographic region of North Mountain. Virtually all the maps on which North Mountain is represented include extensive areas of the neighbouring lowlands as well.

Another problem was that, although the one inch grid was appropriate for the whole area, it would have been useful to have a closer network for the individual physiographic regions. The only solution to the two problems would have been to mark the limits of the physiographic regions on the topographic maps and, using a closer grid, do the height analysis again. The original analysis took so long that it was decided not to repeat the process.

Bearing in mind the reservations expressed above, it is possible to extract some results from the graphs and to offer some tentative explanations. In the Southern New Brunswick Highlands the greatest frequency of heights is at 150-249 feet. Probably the same explanation can be used here as was used to explain the 100-149 maximum for the whole area. In this case, the maximum is at a somewhat greater height because this is a highland region cut by one major river, the Saint John, and many small ones. The second point of note is that between class 650-699 and class 1100-1149 (especially 800-849 to 1100-1149) there is little variation in frequency of points. This distribution is to be expected in a dissected upland region. The comparatively small number of points in the height ranges above 1100-1149 may indicate an accordance of summit levels.

In the case of the Cobequid Mountains and the Antigonish Mountains the "predictable maximum" falls at 200-249 after which there is a steady decrease of frequency with height. However, an exception occurs at 700-749 which may represent an erosional surface.

North Mountain has a maximum of points at 50-99, lower than in the other regions. This is interpreted as being the result of the inclusion of a large number of points on the Annapolis-Cornwallis lowland, a problem already mentioned. Another point of note is that there is little variation between class 250-299 and class 700-749, a situation to be expected in an upland area with one very steep side cut by short, deep valleys and a gentle side which is nevertheless dissected by deep, steep-sided valleys. It is also noticeable that there is a rapid drop in frequency from 700-749 to 750-799; this is attributed to a uniform crest height for the ridge and a certain amount of bevelling.

The Atlantic Uplands show the "predictable maximum" at 100-149, but also less expected maxima at 200-249 and 450-549. This supports the suggestion by Bird and Hare (1956) that rather than being a uniform south-east slope the Atlantic Uplands consist of a series of steps.

Discussion of the Validity of the Peneplain Concept

Some of the implications of the peneplain concept as expounded by Daly and Goldthwait will now be discussed, and conclusions as to the validity of the concept will be made. This involves consideration of three topics: the conditions under which the Triassic sedimentary rocks were deposited; the time of formation of the Fundy syncline; and the scarcity in New Brunswick and Nova Scotia of rocks of post-Triassic but pre-Pleistocene age.

As was noted earlier in this account Klein (1963a) concluded that Triassic deposition took place in a fault trough. He thought that at the time of deposition streams flowed from upthrown highlands to the north and west and from highlands to the south and east. In short, in Klein's opinion a trough existed in the area of the present Bay of Fundy during Triassic times, but it was continental and not occupied by the sea.

The Fundy syncline is responsible for the hook at the Minas Basin end of North Mountain, its southern limb being represented by the north-west dipping Triassic rocks underlying the Annapolis-Cornwallis Valley and by the basalts forming North Mountain (Fig. 46). The northern limb does not have such a clear topographic expression but the syncline has been traced beneath the Bay of Fundy as far south as its mouth (Swift and Lyall 1968a).

Daly (1901, p. 84) referred to North Mountain as "a flat-topped and truncated ridge", the truncation being due to the formation of the Cretaceous peneplain. Therefore, by implication, the syncline must pre-date the peneplain, and, if there is a Cretaceous peneplain, the Fundy syncline must have been formed in the time pèriod between the extrusion of the North Mountain Basalt, and the Cretaceous.

Few references to rocks of post-Triassic but pre-Pleistocene age can be found in the literature. Stephenson (1959) described deposits in the vicinity of the villages of Middle Musquodoboit and Shubenacadie, Nova Scotia that he thought were of Cretaceous age. From his description of the deposits they appear to be of non-marine origin. Later, Stevenson and McGregor (1963) confirmed the Early Cretaceous age of these sediments. Klein (1960, p. 174) introduced some unpublished work by Take under the heading "Discovery of Cretaceous sediments in Nova Scotia", giving the impression that there had been no previous knowledge of the existence of Cretaceous rocks. The lower Cretaceous deposits "discovered" by Take were termed the Shubenacadie Formation, and he suggested that the Bay of Fundy existed at the time of their deposition. Benson (1967) suggested that isolated clay beds in other parts of Nova Scotia, similar to the Cretaceous non-marine clays of the Shubenacadie district, indicate the deposition of more extensive Cretaceous beds. Even so, the known extent of post -Triassic/pre-Pleistocene rocks is limited.

The scarcity of post-Triassic sediments can be explained in one of two ways; either very little post-Triassic sediment was deposited or there was extensive deposition of post-Triassic sediment, virtually all of which has been removed. Each of these possibilities will be considered in connection with the Cretaceous peneplain idea of Daly and Goldthwait and an attempt will be made to show whether or not a Cretaceous peneplain could have been formed across New Brunswick and Nova Scotia.

If there was virtually no post-Triassic sedimentation, would it be possible for a peneplain to form across New Brunswick, the site of the present Bay of Fundy, and Nova Scotia? It has been shown that the Fundy syncline must have been in existence before the peneplain was formed. It would have been impossible for rivers flowing from the north-west to cross the Fundy trough which was in existence by Triassic times and which was later accentuated by the formation of the Fundy syncline. Since the northern limb of the syncline is now beneath the sea, it is difficult to reconstruct what might have happened. However, it is suggested that any rivers flowing south-eastward from New Brunswick were diverted to flow south-westward along the outcrop of softer rock between the upfaulted highlands on the north-west side of the Bay and the westernmost extension of the North Mountain Basalt extrusion. A major river may have flowed south-westward along the axis of the syncline which pitches in that direction. Smaller rivers would have flowed down the flanks of the syncline and into the main river.

Along the southern limb of the syncline, where conditions are clearer because the rocks involved are above sea-level, the drainage pattern was probably such that short streams flowed to the north-west down the flanks of South Mountain and to the south-east down the North Mountain Basalt to a lowland that developed rapidly between the two on the soft rocks of the Annapolis Formation. This lowland is now the Annapolis -Cornwallis Valley which itself would have been drained by a major river flowing to the south-west. This river and its tributaries began undermining the basalt outcrop to begin the formation of what is today North Mountain.

The conclusion the writer reaches from this discussion is that, unless there was extensive post-Triassic deposition, there was no way in which rivers flowing from north-west to south-east across New Brunswick could have crossed the Bay of Fundy area to enter the sea near the present south-east coast of Nova Scotia. Thus, a peneplain could not have developed across New Brunswick and Nova Scotia by Cretaceous time without the existence of a complete post-Triassic cover of rocks.

Having ruled out the possibility of a Cretaceous peneplain developing without extensive deposition of post-Triassic rocks, it is now necessary to consider some of the implications and problems associated with the idea that there was a post-Triassic cover which has been almost completely removed. If it is assumed that there was a post-Triassic rock cover, it must also be assumed that it was thick in parts because in order for there to be a continuous slope from New Brunswick to Nova Scotia the Bay of Fundy trough would have had to be filled. The theory would require that the surface be tilted to the south-east and that rivers flowing to the south-east be initiated on it (consequent streams). These would flow across New Brunswick, the Bay of Fundy, and Nova Scotia and, in so doing, would remove almost the entire post-Triassic cover. The writer finds it difficult to believe that more remnants of the rock cover would not be preserved. In fact, there are probably no "remnants" for Take thought that the Shubenacadie Formation was deposited when the Bay of Fundy was in existence. The Cretaceous rocks are, therefore, isolated deposits rather than part of a cover which used to be over all of New Brunswick and Nova Scotia.

Another line of investigation leads to conflicting results. If the present drainage was superimposed from a post-Triassic cover, it is reasonable to expect a lack of adjustment to structure. There is some evidence of this in the case of the Saint John River between Hartland and Fredericton where the river cuts transversely across the rock outcrops (Fig. 6), but there is negative evidence in the lower part of the Saint John River Valley where the trend of the river is quite definitely determined by structural trends which appear to have existed in pre-Triassic times. However, it could be argued that the structural trends which determine the course of the Saint John and its east bank tributaries between McAlpine and Saint John were so well established that drainage let down from a post-Triassic cover was re-established along the old lines.

If one postulates that across the whole of New Brunswick, the Bay of Fundy, and Nova Scotia a post-Triassic cover existed on which there were south-east flowing rivers entering the sea in the general area of the present south-east coast of Nova Scotia, it is reasonable to search for some evidence of the former paths of these rivers. The question to be answered is whether there is any evidence that the south-east flowing rivers of New Brunswick can be traced across Nova Scotia. Only the major rivers are worth considering because glaciation would probably have destroyed any evidence of the former channels of small streams. The two rivers most likely to yield evidence are the Saint John and the St. Croix.

Since the mouth of the Saint John River is conveniently in line with Digby Gut, it is tempting to speculate that the two are genetically related. Digby Gut is generally acknowledged to be fault determined (Goldthwait, 1924; Cameron, 1956a),but faulting alone, unless it is of the graben type, does not produce a gap through a basalt ridge. Instead it creates a line of weakness which can be exploited by the agents of erosion. It is possible that the erosion that produced Digby Gut was accomplished by a southward extension of the Saint John. There are, however, several other possible explanations: 1) the Bear River flowing in the opposite direction from the Atlantic Uplands may have been responsible for the erosion; 2) tidal scour worked along the line of weakness as sea-level rose to drown the Fundy trough; and 3) glacial erosion was concentrated along the fault zone resulting in the creation of a low-lying area which was later flooded by the sea. A southward extension of the Saint John River is, then, only one of several possible explanations.

In the case of the St. Croix River, it could be lined up with Grand Passage between Long Island and Brier Island except that Grand Manan Island lies in the path of any direct connection. Even if this were not the case much the same alternatives would apply as in the case of the Saint John River and Digby Gut.

It might not be entirely reasonable to expect any evidence to remain in Nova Scotia because, if a peneplain *senso stricto* did develop, the major rivers would be flowing in wide valleys with low interfluves between. Under these circumstances the drainage pattern would be easily deranged by glacial scouring and deposition.

The main conclusion reached from the foregoing argument is that there is no evidence of any extensive post-Triassic deposition, nor is there any clear evidence of superimposition of drainage. This inclines the writer to rule out the possibility of the formation of a peneplain by Cretaceous time by rivers that were initiated on a post-Triassic cover.

Therefore, the Cretaceous peneplain idea of Daly and Goldthwait is not accepted as a valid explanation of topographic conditions in New Brunswick and Nova Scotia. There is, then, no reason to expect any relationship to exist between the erosion surfaces of New Brunswick and those of Nova Scotia. There remains, though, the problem of explaining the origin of the erosion surface that truncates the shales and quartzites of the Atlantic Uplands of Nova Scotia. The idea of exhumed surfaces suggested by Bird and Hare (1956) and Bird (1964) seems to offer the best explanation of this, although their ideas may have been anticipated by Ganong (1904). He noted that, especially in the western part of New Brunswick, the river valleys are largely independent of the geological structure and suggested that they originated on land that was covered by a layer of homogeneous deposits with a slope to the south-east. At that time, the Bay of Fundy may have been a land area and rivers may have flowed across it and Nova Scotia to reach the sea. This appears very similar to the peneplain idea of Daly but the suggestion that the uniform cover was of Carboniferous age eliminates the problems mentioned when considering a Cretaceous peneplain. In the first place, rocks of Carboniferous age are exposed extensively in the Maritimes so the problem of assuming almost complete removal of the cover does not exist. Secondly, postulation of a complete Carboniferous cover would allow for drainage across the Bay of Fundy. Thus, in the writer's opinion, the best explanation for the erosion surface of the Atlantic Uplands is that it was formed by rivers originating on a cover of Carboniferous rock. This is no more than an hypothesis and verification would involve detailed study of the conditions of deposition of the Carboniferous rocks of Nova Scotia, a study which is beyond the scope of this thesis.

Conclusions

Five conclusions may be drawn from map analysis and from discussion of peneplains and erosion surfaces in the Maritime Provinces:

(1) The Bay of Fundy trough was in existence during the Triassic, but was not then below sea-level.

(2) There is no evidence to support the idea of a Cretaceous peneplain. Any planation that took place was pre-Triassic.

(3) The Atlantic Uplands erosion surface consists not of a uniform slope but of a series of plateaux and steps, except in the extreme south-east.

(4) The erosion surface was probably developed by rivers that originated on a cover of rocks of Carboniferous age.

(5) North Mountain is flat-topped in parts, but the truncation was later in date than the formation of the erosion surfaces on the Atlantic Uplands of Nova Scotia and the Southern New Brunswick Highlands.

Evolution from Triassic to Quaternary

Having examined and rejected the peneplain theory, the writer must make an alternative suggestion and, therefore, proposes that the Bay of Fundy and adjacent areas evolved in the following manner from Triassic times to the Quaternary.

A trough was already in existence during the Triassic and deposition by streams flowing to it from the north-west and south-east took place. Extrusion of lava which occurred at times impeded the drainage. At the close of the Triassic or sometime after that, the Fundy syncline, which pitches to the south-west was formed and channelled drainage to the south-west. A major stream which developed along the axis of the syncline flowed to the ocean via the present mouth of the Bay of Fundy and through Northeast Channel, possibly initiating its formation. As Ballard and Sorenson (1968) reported that the channel is cut into Cretaceous sediments this probably occurred in late Cretaceous time. The channel was certainly in existence in pre-Tertiary time because the eroded Cretaceous sediments lie beneath undeformed Tertiary deposits.

On the southern limb of the syncline, streams flowed across the North Mountain Basalt which had been tilted to the north-west towards the present axis of the Bay. Short, swift streams flowed in the opposite direction down the south-east edge of the basalt to join a longitudinal stream which,flowing between the basalt and the older rocks of South Mountain, eroded down to the sandstone and shale below. There was probably a slight south-west tilt to the land so that the stream may have flowed all the way from the Antigonish Highlands to St. Marys Bay along the line of the present Salmon River, Cobequid Bay, Minas Basin, and the Annapolis -Cornwallis Valley. The short streams flowing to the valley from the basalt caused its retreat, thus giving rise to the present scarp-like form of North Mountain.

As the northern limb of the Fundy syncline is below water, its nature is not so well known. Even so it is known that it is not so clearly defined as the southern limb and, as stated earlier, Swift and Lyall (1968a) referred to a half-graben rather than a syncline. No equivalent of Annapolis Valley was formed.

Later in the history of the Bay, relative sea-level rose to drown parts of the Triassic lowlands, in particular the main Fundy trough. At this time, the apex of the Fundy syncline was breached to form Minas Channel, Minas Basin, and Cobequid Bay. In the south, drowning of part of the Triassic lowland led to the formation of St. Marys Bay, with Digby Gut, Petit Passage and Grand Passage being formed at this time also. The faulting which produced the lines of weakness along which these gaps are developed is post-Triassic in age and may have taken place at the same time as the formation of the Fundy syncline. Stream erosion before the rise of sea-level probably played some part in the formation of the gaps. In one instance, Digby Gut, the gap is directly opposite the mouth of a river (Bear River) flowing to the Bay of Fundy from South Mountain. If the basalt extrusion once reached South Mountain, Bear River might have flowed across it, exploiting the weakness caused by the fault. As the river would have been flowing to a base-level which was much lower than the present one, it would have cut deeply into the basalt, producing a narrow gorge. As sea-level rose, and water spread through the gorge, tidal scour would have caused further deepening. Once the basalt ridge had been penetrated, rapid erosion on the soft Triassic sandstone and shale would have led to the formation of Annapolis Basin.

In the case of Petit Passage and Grand Passage, there is no similar river flowing from South Mountain, but in all probability small rivers flowing to the north-west and south-east down the basalt outcrop exploited the lines of weakness caused by the faults. Later, as sea-level rose, the gaps were deepened by tidal scouring.

Not all the faults crossing Digby Neck have produced gaps which penetrate below present sea-level, but one other is very close, that at Sandy Cove. The gap here is located opposite the mouth of the Sissiboo River and the same explanation is suggested for its development as for Digby Gut, except that it is still above sea-level. This might be due to glacial deposition, for there are extensive glacial deposits of various types in the gap.

It is not possible to give a definite date for these events except to say that they must have been post-Triassic and pre-Pleistocene. The reason for the post-Triassic date has already been given, and the pre -Pleistocene date is based on the fact that there is evidence that the Bay of Fundy, North Mountain and Annapolis Valley existed in much their present form during the Pleistocene glaciation.

The sequence of events outlined is an attempt to explain the major topographic features of the study area. As the events occurred in pre -Pleistocene time, much of what has been written is speculative. However, in the writer's opinion, it represents a reasonable explanation of what might have happened. There have been changes which must be explained; for example, the reversal of the drainage direction in part of the Annapolis -Cornwallis Valley, but this can probably be related to the glaciation of the area.

CHAPTER 4

THE PLEISTOCENE: GLACIATION AND DEGLACIATION

Although this is not primarily a study of glaciation and deglaciation, these topics have to be considered in some detail because otherwise it is impossible to interpret evidence of sea-level changes. The early geologists were mainly interested in the bedrock of the area, but much of it is mantled by surficial deposits which could not be ignored. Mapping and classification of the surficial deposits led to speculation about glaciation of the area.

Opinions as to the manner of the glaciation and deglaciation of New Brunswick and Nova Scotia have varied over the years. Dawson (1855a) in *Acadian Geology* explained the unstratified drift or boulder clay of Acadia as the result of frost shattering with the resultant broken-up material being redistributed by floating ice as the land was submerged by the sea. The same explanation was proposed in the second edition of *Acadian Geology* (1868) despite the fact that the modern glacial theory had already met with some support in North America (Dana, 1863). As late as 1893 in *The Canadian Ice Age* he retained the idea that the glaciation of the Maritimes is best explained in terms of local glaciers on the higher areas with icebergs or floating ice on the lower areas during the post-Tertiary submergence.

Belt (1866) was one of the first to abandon the old "drift theory" as an explanation of the distribution of glacial deposits in Nova Scotia.

He argued that the distribution of drift can be explained by postulating a large ice mass developing over the Arctic and moving southward to cover much of eastern North America.

Matthew (1872) also rejected the hypothesis that icebergs and ocean currents produced the phenomena of the "Drift Epoch". He favoured the idea that an ice sheet similar to those now over Greenland and Antarctica used to exist over North America. The general movement of ice over the Maritime Provinces was southward but the movement of bottom ice was governed by the configuration of the land. The western part of the sheet moved from northern Maine down the Atlantic slope to the sea, whereas the eastern part pushed across the "low swell of land separating the Gulf of St. Lawrence from the Bay of Fundy " (Matthew 1872, p. 106). The central part, however, was impeded or nearly arrested by the southern hills of New Brunswick. Some of the ice pushed over the hills or through gaps in them, but the direction of movement conformed to the shape of the land.

Shaler (1874) also talked in terms of an ice sheet. He thought that ice moving along the St. Lawrence estuary was deflected by the peninsula of Nova Scotia so that some of the ice crossed Chignecto Isthmus and flowed into the Bay of Fundy. This ice movement was, in his opinion, partly responsible for the excavation of the Fundy trough.

Later, Honeyman (1879, 1882, 1886), in a series of papers delivered to the Nova Scotia Institute of Science, recorded the southerly transport of boulders from the Cobequid Mountains and North Mountain, his conclusion being that this movement was accomplished by ice that originated beyond Nova Scotia.

Chalmers (1885) was of the opinion that the whole of New Brunswick was covered by ice in the early part of the Quaternary, but that most evidence is in favour of a number of small local glaciers whose movement was influenced by the shape of the land; for example, local glaciers probably formed on higher ground and crept down into Shepody Bay and Cumberland Basin, thus covering Chignecto Isthmus. In 1890 he claimed that the Saint John Valley with the lake basins to the east of it and the various tributary valleys were filled by a confluent mass of ice which moved towards the coast. By 1895 he was even more strongly in favour of the theory of local glaciers and he stated: "The theory of local glaciers upon the higher grounds and floating ice on the lower coastal districts proposed by me in 1885 and 1886 as a working hypothesis for the explanation of the Pleistocene glacial phenomena of New Brunswick and south-eastern Quebec may now be considered, with certain exceptions and modifications, as established " (Chalmers 1895, pp. 87M-88M).

He dealt at some length with the various local glaciers that he claimed to be able to distinguish. Those affecting New Brunswick were: 1) the Northumberland glacier which had its source in the highlands of central New Brunswick and extended eastward from them across the Northumberland Strait and over a part of Prince Edward Island; 2) the Saint John Valley glacier which originated in the highlands of northern Maine, the eastern townships of Quebec, and north-western New Brunswick; it was joined by large tributary glaciers but probably did not over-ride the crystalline plateau lying to the east of the mouth of the Saint John River; 3) a local ice mass which accumulated on the crystalline plateau east of Saint John and flowed out towards the open waters of the Bay of Fundy; 4) the Chignecto glacier which existed on the Chignecto Isthmus and in the two arms of Chignecto Bay, extending as far south as Cape Enragé and Apple River; the general movement of the glacier was southward into the open waters of the Bay of Fundy.

In the case of Nova Scotia, he was convinced that no ice had reached the province from the mainland; that is, the Bay of Fundy was not crossed by land ice from southern New Brunswick, nor was Nova Scotia glaciated by ice from the north or north-east. He believed that ice gathered on the summit of the Cobequids and moved both northward and southward, and that on the south side a large local glacier moved westward from Minas Basin into the Bay of Fundy. In the east, a local glacier flowed off the Cobequids across Cobequid Bay and then south to Halifax. In his opinion, over the main part of Nova Scotia ice accumulated locally and then flowed outward. Ice accumulating on South Mountain, for instance, moved across Annapolis Valley, over North Mountain, and into the Bay of Fundy.

Chalmers' opinion that ice did not cross the Bay of Fundy is in contrast to that of other workers. Prest (1895-1896, p. 168) for example stated: "That there has been a time when a continental glacier ploughed its way across the Bay of Fundy and the Province of Nova Scotia, seems to be beyond doubt." Bailey (1894, 1897, 1898, 1910) also thought that ice moved across the Bay of Fundy bringing debris from New Brunswick and depositing it in Nova Scotia. When the main ice disappeared, local centres of ice came into existence and movement took place in the opposite direction, for example in Annapolis Valley.

The idea of a reverse movement was supported by Walker and Parsons (1923) who concluded that, although there may have been a general continental glaciation of Nova Scotia, "there has certainly, as a last act, been a northward movement from the highlands of north-eastern Lunenburg, which is [sic] now over 800 feet above sea-level, across the Annapolis Valley and the North Mountain to the Bay of Fundy at Hampton " (Walker and Parsons 1923, p. 8).

Goldthwait (1924) extended the ice sheet further south than previous writers, contending that there is proof that it filled the Bay of Fundy and covered the banks off the south coast of Nova Scotia. He thought two lobes of ice affected Nova Scotia; one, which he termed the New Brunswick Lobe, flowed eastward from New Brunswick reaching eastern Prince Edward Island; the other, the so-called Acadian Lobe, covered the northern part of Nova Scotia reaching the head of the Bay of Fundy. The two overlapped considerably but they do not represent separate epochs. He ruled out the idea of local ice caps on the Atlantic Uplands of Nova Scotia, on the Cobequid Mountains, and on Cape Breton Island.

In 1930, Shepard, discussing the origin of the steep north-western shore of the Bay of Fundy, argued the case for the steepness being due to glaciation. He referred to a Fundian Glacier which occupied the main part of the Bay of Fundy and was formed by lobes from Minas Basin and from Chignecto Bay and by tributaries from the Saint John Valley, the Passamaquoddy Bay Lowland, and St. Marys Bay (Fig. 47).

Alcock (1948, 1949) believed that the whole province of New Brunswick was overrun by ice moving in a southerly direction; that is, by the Labrador ice sheet. He thought that local glaciers existed over the central highlands of New Brunswick during two separate time periods: 1) they accumulated over the highlands before the main glaciation and were overrun by the Labrador ice sheet as it moved down from the north; and 2) after the dissipation of the main ice sheet, local ice centres remained. He also made a case for movements of ice eastward from the mountains of Maine. This ice crossed part of the Bay of Fundy and overrode Grand Manan Island.

MacNeil and Purdie (1950-1951) talked of the Labrador ice sheet covering Nova Scotia and then the development of a local ice cap in the area of Shell Bog Lake on the Atlantic Uplands. This local ice spread northward into the Annapolis-Cornwallis Valley and up onto North Mountain. MacNeil (1951b) thought it reached the Bay of Fundy.

Flint (1951) reviewed the evidence for local centres of glacial outflow. In the case of the Central Highlands of New Brunswick, he concluded that local ice preceded the Laurentide ice sheet and later merged with it, and that striations within the area record local ice movement after the Laurentide ice sheet disappeared. In the case of the Atlantic Uplands he concluded that "despite the low altitude of the district the concept of a local center of radial outflow seems inherently probable " (Flint 1951, p. 30). The evidence for and against a centre of outflow from the Cobequid Mountains was reviewed but no conclusion was reached.

Lee (1953), however, showed three highland centres of outflow in New Brunswick and Nova Scotia: 1) the Central Highlands of New Brunswick; 2) the Cobequid Mountains; and 3) South Mountain. He thought that, at the maximum stage of glaciation, the ice front was well south of Nova Scotia and that as the ice disappeared it split into two sections along the line of the Bay of Fundy so that the Bay was partially open while the surrounding land areas were covered by ice. At a later stage there were local remnants on higher areas which accounted for movement in the opposite direction.

Prest *in* C. H. Stockwell (1957) stated that Nova Scotia was affected by ice that moved in a generally south-east direction from the highlands of New Brunswick across the Bay of Fundy and across the greater part of Nova Scotia. In addition to this, on the basis of south-trending striae in the north central part of Nova Scotia, he argued the case for glacier ice moving eastward along Northumberland Strait, encountering ice moving out from Cape Breton Island and being deflected to the south. Over New Brunswick the Laurentide ice covered the whole area at the Wisconsin maximum, moving in a south to south-easterly direction with strong movements down the Madawaska - Saint John Valley. Later there was an ice cap on the New Brunswick Highlands with radial outflow.

Hughes (1957), working in the Shubenacadie area, concluded that Cobequid Bay and the lower reaches of the Shubenacadie River became ice free before the interior of Nova Scotia. An ice mass was detached from the main ice sheet and this inert ice mass south of Cobequid Bay then wasted *in situ* by marginal melting and surface ablation.

Hickox (1962b), working in Annapolis Valley, talked of an ice cap centred along the axis of the peninsula of Nova Scotia. This thickened sufficiently to flow out radially and the outflow advanced northward at least 15 miles to reach the Bay of Fundy shore.

Grant (1963) suggested that ice moved from a Cape Breton centre westward onto Prince Edward Island and south-westward along the Atlantic coast of Nova Scotia. New Brunswick ice moved across the Bay of Fundy and also eastward across Northumberland Strait. The Cape Breton ice and the New Brunswick ice met to give a series of currents flowing in a general southerly direction: 1) the Chignecto glacier; 2) a southerly flow at right angles to the Cobequid Mountains; and 3) a south-south-westerly flow from George Bay.

Borns (1965, 1968) tentatively correlated the last ice sheet to cover northern Nova Scotia with that which constructed an end moraine system extending from Cherryfield, Maine, through Eastport, Maine, to Saint John, New Brunswick. On the basis of a radiocarbon date [I(GSC)-7], he concluded that ice was probably present along the position of the moraine 12,325± 500 years ago. The ice sheet receded northward leaving end (recessional) moraines at Saint John, Woodstock, and Grand Falls, New Brunswick and at St. Antonin, Quebec, reaching the St. Antonin/Trois -Pistoles area 12,720± 170 years ago. He favoured the idea that residual ice remained over South Mountain and that later an ice cap developed with ice flowing out radially. The Cobequids were not a centre of radial outflow after the dissipation of the last ice sheet but have supported a more extensive snow cover than they do now.

Take (1965) thought that deglaciation began in Nova Scotia at least 15,000 years ago and that the continental ice sheet had receded from all areas by 13,000 BP with a readvance across North Mountain taking place at least 14,500 years ago. He thought that the possible or proposed ice caps on the Nova Scotia mainland were in contact with the main continental ice sheet during most of their existence and that they also disappeared by 13,000 BP.

On the basis of his work on the surficial deposits of south-western New Brunswick, Gadd (1968, 1969, 1970a) concluded that at the maximum of the last glaciation continental ice moved across south-west New Brunswick towards the south-east and south, the southerly trends being characteristic of the area east of Fredericton and St. George and the south-easterly trends being characteristic of the area to the west. Large moraines and associated outwash deposits between St. George and Saint John were formed by the continental ice sheet which overrode the hills in the coastal part of the area. After formation of the coastal morainic systems, an ice lobe probably occupied the St. Croix Valley and Passamaquoddy Bay. The latter was ice free and the ice margin had retreated from the area north -west of St. Stephen some 12,400 years ago. After this there was lobate ice margin retreat to the north-west.

In recent years, much attention has been given to evidence of glaciation in areas which are now below sea-level. Rvachev (1965) believed that Georges Bank lies close to the southern boundary of the Quaternary ice and that ice spreading onto the shelf probably occupied only the northern part of the Bank. He thought this ice probably thrust forward in the form of large tongues, one of which pushed along Northeast Channel. This accounts for the smaller depth at the mouth of the Channel as compared to its depth in the Gulf of Maine, glacier ploughing having taken place in the Gulf and the accumulation of moraine having occurred at the edge of the glacier.

Emery and Uchupi (1965) and Uchupi (1966a, 1966b) were of much the same opinion, suggesting that ice occupied the Gulf of Maine and terminated against the landward slope of Georges Bank. The ice deepened and widened the Gulf and a valley glacier moved through Northeast Channel, the Gulf of Maine being ice free by 11,000 years ago.

The Glacial Map of Canada (Prest, Grant, and Rampton, 1968) shows part of Browns Bank and the Scotian Shelf as being mantled by moraine and

King (1969) recognized a submarine end moraine system on the Scotian Shelf. He concluded that the Scotian Shelf was completely glaciated during the Pleistocene, but that it is not clear whether the last ice sheet extended to the edge of the continental shelf. Pratt and Schlee (1969) believed that the maximum limit of the last glaciation was probably along a line extending across Great South Channel, northern Georges Bank,and at least to the edge of the Scotian Shelf. The ice margin was probably a subaerial one on Georges Bank, but on the Scotian Shelf the margin may have bordered directly on the ocean. A lobe of ice pushed into Great South Channel, and Northeast Channel was "an active thoroughfare for glaciers."

Additional information is forthcoming from the Bay of Fundy itself (Swift and Lyall, 1968b). It is argued that closed depressions in Chignecto Bay, Minas Passage and at the mouth of the Bay of Fundy cannot be attributed solely to modern tidal scour and, while they may have originated as subaerial river valleys, the owe their present configuration to ice action. Their orientation indicates that Fundy at some point underwent longitudinal glaciation. Evidence in favour of transverse glaciation is also quoted. It takes the form of a moraine near Port Lorne, Nova Scotia whose orientation indicates ice movement at right angles to the shore, and a bedrock river channel running roughly parallel to the axis of the Bay which is interpreted as an ice marginal feature developed along the front of an ice lobe.

Opinions with respect to the glaciation and deglaciation of the Maritimes are continually changing, for instance Prest (personal communication) thought that ice moved across the Chignecto Isthmus. He first ascribed this to lobing of New Brunswick ice moving east along Northumberland

Strait, but in 1968 thought that this ice flowed southward from Prince Edward Island and that, as sea-level rose, long leads were formed in the front of the ice. After re-interpretation of linear glacial features, he concluded that Laurentide ice was not as active in the Maritime Provinces as is commonly believed and that the growth of Appalachian glaciers during the build-up of the last continental ice sheet effectively barred Laurentide ice from some parts of the region (Prest and Grant, 1969). Rising sea-level, from about 18,000 to perhaps 10,000 years ago, imposed a pattern of radial flow from several land areas that was maintained by local upland nourishment under the maritime climatic regime. The map, *Retreat of Wisconsin and Recent Ice in North America* (Prest, 1969), showed retreat of the ice sheet from the Scotian Shelf by about 18,000 years ago, to the mouth of the Bay of Fundy by 14,000, most of the Bay being uncovered by 13,200. Ice remained over southern Nova Scotia, northern Nova Scotia and Prince Edward Island,and the northern highlands of New Brunswick until considerably later.

From this summary of ideas on the glaciation of the Maritime Provinces, it is possible to distinguish a number of points of dispute and certain underlying trends in thinking on the subject. In the early days, the main dispute was between those who explained the "drift" as being deposited by icebergs and those who thought it was deposited by land ice rather than floating ice. Later there was considerable controversy as to whether the glaciation had been carried out by local glaciers or by an ice sheet sweeping down from the north. After initial uncertainty, the ice sheet idea, with some modifications, gained supremacy and the crossing of at least New Brunswick by the Labradorean Ice Sheet was accepted as fact. But as late as 1920 there was opposition to this (Coleman 1920).

Very recently, Prest and Grant (1969) stressed the importance of the development of local (Appalachian) ice caps, thus reopening the argument.

Another controversy for many years was whether or not ice from New Brunswick had crossed the Bay of Fundy and Nova Scotia. Modern thinking is that it did and recent work shows fairly conclusively that, at its maximum extent, it reached the Scotian Shelf well south of Nova Scotia. A related question is whether there was movement of ice across the Chignecto Isthmus and down the Bay of Fundy. Recent submarine data (Swift and Lyall, 1968b) seem to point to both north-west/south-east movement across the Bay and longitudinal movement along the axis of the Bay causing some deepening, although there is little to support the contention that the Bay of Fundy is primarily of glacial origin.

The existence of remnant ice caps after the dissipation of the main ice sheet is another question which has received much attention in the literature. An ice cap over the Southern Uplands of Nova Scotia which spread north-west and resulted in the movement of glacial debris across Annapolis Valley and North Mountain seems to meet with general acceptance. A second centre located over the highlands of northern New Brunswick has some support, but a third, Cobequid Mountain, centre is much more doubtful, although supported by Prest and Grant (1969).

Within recent years much attention has been focussed on evidence to be found below present sea-level. As a result, the nature and possible mode of origin of the bottom deposits in the Bay of Fundy, Gulf of Maine, Northeast Channel, Georges Bank, Browns Bank, and the Scotian Shelf are becoming much clearer. Attention is also being focussed on the mode of retreat and/or dissipation of the ice sheet that once covered New Brunswick and Nova Scotia. This is a welcome trend from the point of view of the writer because interpretation of indicators of sea-level changes is dependant on knowledge of the land/ice/water contact at stages in the past.

Since the publication of Libby's book, *Radiocarbon Dating* (1952), it has been possible to obtain precise dates for late Quaternary events with some dates available for New Brunswick and Nova Scotia.

CHAPTER 5

QUATERNARY CHANGES OF SEA-LEVEL

Complicating Factors

Introductory Remarks

The Quaternary Period is the best known of the geological periods, not only in the region under discussion in this thesis but throughout the world. This is due partly to availability to the geomorphologist of a range of techniques which cannot be applied to earlier geological periods. However, the increased range of techniques available and the increased factual knowledge do not necessarily make the picture any clearer. In fact, the wealth of material available frequently makes interpretation of events more difficult since there are more facts to fit into the overall framework. Moreover, the range of possible explanations is greater, and it is often difficult to decide which is the best of the alternatives. Also, it is easy to arrive at conclusions that appear to contradict one another.

The emphasis of this thesis is on changes of sea-level around the Bay of Fundy, in particular on the amount of change, the causes of any changes and, where possible, the time at which they took place. Such a study would be complex in any part of the world, but the situation is more complicated in the Bay of Fundy than in other areas, principally because of the great tidal range experienced by the Bay. As has been explained, in all parts of the Bay, a great difference between high-tide level and low-tide level exists, but there is no certainty that this was the situation throughout the Quaternary. The tidal range is dependent on size, depth and shape of the Bay, all of which have changed during the Quaternary Period as a result of glaciation and the changing level of the sea relative to the land. Almost certainly then the tidal range has changed, complicating the interpretation of indicators of previous sea -levels. An added complication is that the relationship between contemporary morphological features and various levels of the tide is not fully understood. There have been few studies of this nature anywhere in the world and, as far as the writer knows, none in the Bay of Fundy.

Glaciation complicates the picture in that it was directly and indirectly responsible for alterations in size, shape, and depth of the Bay during the Quaternary and, consequently, for changes in tidal range. Moreover, the existence of an ice cover over the Fundy coast when relative sea-level was higher than at present may have prevented the preservation of any record of higher sea-level.

It is also necessary to consider the question of eustatic (world -wide) changes of sea-level which were superimposed on local changes and may, in fact, have cancelled out local changes in some instances and amplified them in others.

Causes of the Present Tidal Range

The tides of the Bay of Fundy were mentioned briefly in Chapter 2, but the approach was purely descriptive and no attempt was made at that time to explain the great tidal range. The fact that the Bay of Fundy has the greatest known tidal range in the world is a result of the relationship between the tide producing forces and the Bay's size, depth, and shape.

In order to know whether tidal conditions in the past were different from those of today, it is necessary to understand, at least in general terms, the reasons for the present very large tidal range. It has long been known that this is due in part to the narrowing of the Bay in a north-easterly direction; for example, Chalmers (1895, p. 16M) wrote: "The greater rise of the tides in the upper parts of this bay is attributed to its narrowing funnel form, and its shallowing bottom cooping up the tidal wave as it advances up the bay." Unfortunately, in the literature there was a tendency to overemphasize this cause, but it can hardly be held to account for the smaller, but nevertheless large, tides at the mouth of the Bay. The size, shape,and depth of the Bay as a whole have to be considered.

The *Atlantic Coast Tide and Current Tables* (1966, p. 2) describe the tides of the Bay of Fundy as follows:

> The tides of the Gulf of Maine are ... propogated into the Bay of Fundy. The decrease in cross-section of this bay, toward its head, is one factor which increases the range of both semidiurnal and diurnal tides. The most important factors are, however, the length and depth of the Bay, for it is these dimensions which determine its natural period of oscillation.

Something more specific than these general descriptions is required. Two types of wave motion are considered in tidal theory: the progressive wave,and the standing wave or standing oscillation which is applicable mainly to gulfs and basins (King 1962, pp. 162-166). The period of the standing oscillation can be related to the dimensions of any basin in

question using Merian's formula: $T = \frac{4L}{\sqrt{gd}}$, where T = natural period,
L = length of basin,
g = force of gravity, and
d = depth.

Thus, the period depends on the length and the depth of the basin, each basin having its own natural period of oscillation which depends on these dimensions. If this natural period of oscillation is the same as that of the tide-producing forces, the basin will respond to the appropriate tidal period and a state of resonance will be set up. The critical length for a gulf is given by the formula $L = \frac{1}{4} T \sqrt{gd}$; that is, the critical length varies with the period, T, and the depth, d.

By King's estimation (1962, p. 175), the Bay of Fundy averages 225 feet in depth and its critical length is 160 miles which agrees closely with the measured length of 162 miles. The natural period of oscillation of the Bay is 6.29 hours which almost fulfils the condition for resonance to occur; that is, for movements caused by the standing wave to be reinforced by other tidal forces, in this case the 12.42-hour lunar semi -diurnal tidal forces.

Additional factors in the tidal situation are the shallowing of water towards the head of the Bay and its narrowing and bifurcation at Cape Chignecto. These result in an increase in tidal range towards the extremities of the Bay and the development of the bores on the Petitcodiac, Maccan, and Salmon rivers.

Tidal range is greater on the south side of the Bay than on the north, and greater in Minas Basin and Cobequid Bay than in Cumberland Basin and Shepody Bay. The flooding tide moves mainly up the south side of the Bay while the ebbing tide moves mainly along the northern side. This results in a counter-clockwise system of residual tidal currents. This "Kelvin Wave" is attributed to the rotation of the earth.

Detailed explanations of the tidal phenomena of the Bay of Fundy are available (Redfield, 1950; Rao, 1968), but even from the foregoing relatively simplified explanation it can be seen that any change in the size, depth, or shape of the Bay would produce an alteration in the tidal conditions. Change took place in all three during the Quaternary; therefore, any indicators of past higher or lower sea-levels may have been formed under different tidal conditions from those that exist today.

Glaciation

Effect on tidal range

It is generally accepted that the earth's crust is depressed by the weight of ice as it accumulates and that, if an ice cap covers a region for a sufficient length of time, isostatic equilibrium will be achieved. When the ice dissipates, there is a rebound of the crust to the position it occupied before depression. It has been shown that ice covered the whole of the Maritime Provinces during the Quaternary; it is reasonable, therefore, to suppose that there was depression of the crust. A general 50-foot lowering of the land relative to the sea would result in the drowning of Chignecto Isthmus and the Bay of Fundy would no longer exist as a bay; thus the conditions that account for the present large tidal range would be eliminated. However, if the Isthmus were plugged by ice when relative sea-level was higher, the Bay would have a form similar in general outline to that which it has today, except that the northern shore would be a water/ice contact rather than a water/land contact. The tidal situation might, therefore, have been similar to that which now exists.

The exact sequence of events during the withdrawal, dispersion, or dissipation of the ice is not fully understood. The main problem from the point of view of this study is to decide when Chignecto Isthmus became free of ice because, theoretically at least, marine features now above sea-level may have been formed under three different situations: 1) when Chignecto Isthmus was plugged by ice; 2) when Chignecto Isthmus was free of ice and below sea-level; or 3) when Chignecto Isthmus was free of ice but above sea-level.

Whatever the sequence of events during deglaciation, it is likely that, for at least part of the time between deglaciation and the present, the tidal range was different from what it is now. Interpretation of morphological indications of former sea-levels has to be made with this fact in mind.

Effects on the preservation of evidence of former sea-levels

Even if there are no complications due to changes of tidal range, it is not correct to assume that the difference between present sea-level and the highest marine feature represents the maximum difference in relative level that occurred. According to Andrews (1968a):

$S = Ur+Up+Urr$, where S is the amount of depression attributed to plastic deformation resulting from the weight of ice;

Ur is the amount of restrained rebound; that is, the amount of uplift accomplished prior to deglaciation;

Up is postglacial uplift; and

Urr is the residual depression existing at present.

There will be no morphological expression of Ur or Urr (above present sea-level) and any attempt to assign values to them has to be done on the basis of predicted uplift curves (Andrews, 1968b). In the case of Up it is reasonable to expect that, given favourable conditions, some evidence of former sea-levels will be preserved. It should be remembered, however, that post-glacial uplift is the sum of the present elevation of the marine limit and the appropriate eustatic sea-level correction.

So, even under the best possible conditions, morphological evidence of former sea-levels taken from a single locality is an incomplete indicator of any sea-level change that has taken place.

Problems of Interpretation resulting from Changes of Tidal Range

Few studies have been made of the relationship between coastal morphological features and the height of contemporary sea-level, not even in relatively simple areas,much less in areas such as the Bay of Fundy with its complicating factor of a large tidal range. There is even some doubt as to what is meant by "contemporary sea-level". It might seem adequate to take it as mean sea-level calculated over a long period of time, but there are two objections to this: 1) few stations have records extending over long periods of time, the only one on the Fundy coast being Saint John which in 1969 had records for a period of 36 years; and 2) it is thought that sea-level is undergoing slow eustatic change (Gutenberg, 1941; Marmer, 1949; Donn and Shaw, 1963). Calculation of an average from many years' records obscures these changes. Therefore, the definition given is of doubtful validity and usefulness from the point of view of the geomorphologist. Moreover, mean sea-level is an abstract idea and has no morphological expression unless the sea is virtually tideless. If there is no morphological expression of contemporary mean sea-level - assuming it is possible to arrive at a satisfactory definition - it is not reasonable to expect to find any morphological expression of past mean sea-levels unless the sea in the past was tideless. For these reasons it is impossible to make measurements of recent and fossil beaches refer to mean sea-level as a standard, as suggested by Zeuner (1952).

If mean sea-level has no morphological expression, perhaps high-tide level and low-tide level are more significant when studying changes of sea-level. Any morphological expression of high-tide level would be more useful than an expression of low-tide level because the feature representing contemporary high-tide level would be above water level most of the time and therefore normally available as a base point for comparison with fossil features of the same nature now above sea-level. Conversely, any morphological expression of contemporary low-tide level will be below water most of the time and therefore not easily available for comparative purposes unless elaborate equipment is available.

It is worth considering, then, whether there is any easily identifiable morphological expression of high-tide level. Zeuner (1958) listed a number of indications of erosional origin among them being the junction of the wave-cut platform and the cliff. As this is said to be at high -water mark or a little higher it gives an approximate value for high-water level. In addition, at the platform/cliff junction there is sometimes an undercut or notch which is thought to be the best indicator of high-water mark.

Although the sea can be seen to reach above the base of the cliff at times of high-tide, it is impossible to judge by purely visual methods just how far above the cliff the water reaches. As a result, along some of the cliffed coast there is no indicator of contemporary high-tide level. However, in some parts of the coast the notch mentioned by Zeuner (1958) is obvious, for example between Ogilvie and Morden, Nova Scotia (Fig. 36), and in the case of some stacks, for example at Hopewell Cape, New Brunswick, there is a marked undercut (Fig. 20). The notch and the undercut probably represent high-tide level, but systematic study of the relationship is necessary before this can be stated with certainty.

The situation that exists, then, is that along the cliffed coast of the Bay of Fundy there is no definite morphological evidence of high -tide level. If this is the situation with respect to contemporary coastal morphology, it is not reasonable to expect any morphological indicator of past high-tide levels. Along parts of the Fundy coast a raised wave-cut platform and cliff are preserved, but it is not possible to say with certainty what tidal level is represented by the junction of the two and nowhere along the entire coast of the Bay was the writer able to locate an example of an uplifted notch. All that can be said with certainty is that on occasions the sea reached a level as high as the platform/cliff contact. Moreover, it is not to be expected that after a period of some thousands of years the raised platform and cliff would retain their original form. It is difficult, therefore, to locate the junction between the two (Fig. 48), with the result that the recorded altitude of the contact can vary by several feet depending on where the observer thinks it is located. Weathering, mass wasting, and subaerial erosion since the time of uplift probably account for the elimination of wave-cut notches.

In fact, using only morphological features of erosional origin, the best that can be done at any one locality is to compare: a) the highest indication of present sea-level, which is not necessarily the highest level reached by the sea; and b) the highest indication of a past sea-level, which is not necessarily the highest level that was reached by the sea. Errors of many feet can be made if attempts are then made to calculate high-tide level and more particularly mean sea-level.

The following examples are theoretical but illustrate the errors that can be made:

Example 1. - Along a contemporary cliff the sea at high-tide reaches 4 feet above the base of the cliff, but a research worker takes the platform/cliff contact as being indicative of high-tide level. He is, therefore, making an error of minus 4 feet. Behind the contemporary cliff there is a raised wave-cut platform and cliff with a clearly-defined junction between the two located at a height of 50 feet above the present contact. If it is assumed that conditions during their formation were basically the same as those in existence today and that the sea reached 4 feet above the platform/cliff junction which was, however, taken as high-tide level, the error would again be minus 4 feet. Comparison of the height of the two platform/cliff junctions would give the relative change of high-tide level because the errors were the same in each case and "like things" have been compared. However, the assumptions made with respect to the uplifted platform and cliff are hardly valid. As has already been suggested, the platform/cliff contact is not likely to be sharp and faulty location of it can result in errors of several feet. Moreover, with sea-level approximately 50 feet higher than at present, the dimensions of the Bay would have been different and, consequently, the tidal range would not have been the same as it is now. Finally, with change in sea-level, coastal configuration would be altered, thus altering the fetch and the height reached by waves. For these reasons, it is extremely unlikely that the error made for the uplifted platform/cliff junction would be the same as that made for the contemporary one. Therefore, the comparison would not be between "like things" and the result would not give a true indication of relative change of high-tide level.

Example 2 (Fig. 49). - The errors made in trying to calculate and compare high-tide levels on the basis of morphological evidence are not as great as those likely to occur if attempts are made to calculate and compare mean sea-levels. For the sake of simplicity, it is assumed that the platform/cliff junction does represent high-tide level and that a contemporary junction and a fossil junction, at a height of 75 feet above the present one, are clearly defined and easily located. If the area in question has a tidal range of 30 feet, not excessive in the Bay of Fundy, mean sea-level can be taken as 15 feet below the platform/cliff junction. Assuming no change in tidal range, the old mean sea-level was 15 feet below the uplifted junction. Mean sea-level has therefore fallen 75 feet. However, if the tidal range was different - and it probably was - then this conclusion is entirely erroneous. If it is assumed that the tidal range was six feet at the time of formation of the uplifted features, mean sea-level would have been 75-3=72 feet. Present mean sea-level is -15 feet if the platform/cliff junction is taken as zero. Therefore, the change of mean sea -level amounts to 72+15=87 feet, not 75 feet as previously calculated. The possible error resulting from this one factor is 12 feet and more extreme cases could be cited.

Because of the probability of the occurrence of this type of error the writer believes that it is more meaningful to compare "like things" which have some morphological expression than to attempt to compare calculated values, and that any statement about changes of sea-level in the Bay of Fundy should be accompanied by a statement of the possible sources of error.

At this point it is worth considering whether there are any depositional features of use in determining sea-level changes. A "beach" as defined in Chapter 2 is a deposit that can extend from mean low-water mark to the effective limit of storm waves; that is, to the extreme high-water level. This means that it may have a large vertical range, especially along coasts with large tides. Moreover, beach is not usually continuous over the whole area mentioned in the definition and often occurs as isolated pockets which may be near high-tide level, low-tide level, or somewhere in between. For these reasons, its variability of height with respect to the sea and its discontinuous nature, it is not a reliable or precise indicator of sea-level change. However, again it is possible and desirable to compare "like things". The highest point of a contemporary beach indicates that, on occasions, the sea reaches that level. Similarly, the highest point of a beach now well above sea-level indicates that, on occasions in the past, the sea reached that level. If measurements are made of the difference of height between the two, statements should make no implication that comparison is being made between two high-tide levels, two low-tide levels, or two mean-tide levels. Statements may imply that comparisons are being made of the results of extreme conditions and not those of "normal" conditions. It has to be borne in mind that the height to which beach material can be thrown depends largely on the fetch - assuming there is no basic change in wind conditions - and that, with a higher sea -level, coastal configuration and consequently fetch may have been altered. Delta surfaces provide another possible indicator, but again there is the problem of relating an uplifted delta surface which may have been formed when the tidal range was only a few feet with a surface formed under the present tidal conditions.

During fieldwork, the writer often found it convenient to use the debris line (composed of logs, bottles, and other flotsam and jetsam) at the back of a beach or an area of tidal marsh as a base point from which to take measurements to marine features above sea-level. It must be realized, of course, that the debris line represents an extreme high-water level and that there is no fossil equivalent. However, provided it is realized and stated that "like things" are not being compared, the method is valid.

Other depositional features that might be used are spits and bars, but their relationship to tide levels varies. Some parts are below even low-tide level, while other parts may be many feet above "normal" high-tide level. Spits and bars are not, therefore, good morphological indicators for precise determination of sea-level changes. However, as with other features discussed, they can be used as indicators of former sea-levels providing the possible errors are recognized and great accuracy is not claimed.

So far only indicators of emergence have been mentioned. In the case of submergence vegetation is a good indicator in that it is sensitive to changes in salinity, an increase in which rapidly kills off some plants. Around the Bay of Fundy are numerous examples of "submerged forests". Accurate levelling of the dead tree stumps and the dating of them permits comparison of present sea-level with that in existence when the trees were alive. Grant (1970) used the hydrographic datum termed "Higher High Water at large tides" which is the tidal level attained fourteen times per year arranged over 20 years. "Botanically, it corresponds to the level at which trees are killed by salt water and which is marked by sedge plants (cattails, reed, and salt-water bulrush) spanning the transition between truly marine and freshwater bogs and upland vegetation " (Grant 1970, p. 677). By radiocarbon dating of this datum where it is now below high-tide, he was able to work out the rise of sea-level in the Bay of Fundy over the last 4500 years. However, it will be shown later that changes in tidal range cannot be ignored in the case of submergence any more than in the case of emergence.

The conclusions drawn from the foregoing discussion are: 1) The relationship between contemporary morphological features, both erosional and depositional, and sea-level at various stages of the tide in the Bay of Fundy is not fully understood; 2) This being the case, it is not possible to relate uplifted features to a specific tide level; 3) The situation is complicated by the fact that the tidal range was almost certainly different from that of today when uplifted marine features were formed; 4) Great accuracy should not be claimed for statements about changes of sea-level determined from morphological features. There should be a clear statement of what is being compared and sources of error should be mentioned; 5) In the case of submergence, vegetation is a useful indicator but changes in tidal range cannot be ignored; 6) Any light that can be thrown on the pattern of glacial retreat from the area and the development of the present outline of the Bay of Fundy will simplify interpretation of indicators of sea-level change.

There are two methods of trying to solve the problem of what the tidal range was when an indicator of higher sea-level was formed. The first is to attempt to reconstruct the most likely outline of the Bay at that time and from this to predict a likely tidal range. In this connection the geomorphological history of Chignecto Isthmus is most important. The second line of investigation is to observe deposits and other marine indicators now found above sea-level and to determine whether they have characteristics which result from formation in areas with large tidal ranges. There are problems, the main one being the relative scarcity of studies of deposition in areas with a large tidal range. The situation is changing, however, and during the last few years some detailed studies have been carried out in the Minas Basin - Cobequid Bay area (Klein, 1963b, 1964, 1966; Swift and McMullen, 1968; Swift, McMullen and Lyall, 1967; Swift *et al.*, 1969).

The second method was used by Swift and Borns (1967a, 1967b) in their investigation of a raised fluviomarine outwash terrace on the north shore of Minas Basin. They concluded that the tidal range in Minas Basin was much less when the terrace was deposited that it is now. Hickox (1958), on the other hand, reported an uplifted cliff and wave-cut platform between Harbourville and Port George, Nova Scotia, the cliff/platform junction being 65 feet above the modern equivalent. Because the emerged shore and the modern shore are similar in many respects, he arrived at the conclusion that the tidal situation was much the same in both cases. However, if the formula, $T = \frac{4L}{\sqrt{gd}}$, is used and the appropriate measurements at the time of formation of the Harbourville - Port George "strandline" are substituted in the formula, although L would be unknown and would have to be assumed to be the same as it is today, d would definitely be greater than at present (because sea-level was 65 feet higher than it is now). T would, therefore, have a different value and conditions for resonance would probably not exist: consequently tidal range would be less than it is now. Much depends on the value of L which points to the need to determine the outline of the Bay in late-glacial and post-glacial time.

Eustatic Changes of Sea-Level

Eustatic changes of sea-level are world-wide and attempts are made to distinguish between them and isostatic changes which are more localized, although in fact the two cannot be separated. There are three principal causes of eustatic changes: 1) major sedimentation (sedimento-eustasy); 2) earth movements that affect the shape of the ocean basins (tectono-eustasy); and 3) the build-up and melting of ice sheets (glacio-eustasy). There are other minor causes such as change in the volume of water resulting from a change in temperature (Fairbridge, 1961).

Glacio-eustatic changes are closely associated with glacio-isostatic changes, the latter resulting from depression or rebound of the earth's crust with the addition or removal of ice. It has been pointed out that rebound begins once the ice begins to thin and not when it has disappeared. A fall of relative sea-level results from the rebound. However, ice thinning adds water to the oceans which causes them to rise. Therefore, in any area, such as New Brunswick and Nova Scotia, covered by ice during the Quaternary there will be an interplay of sea-level changes due to depression and rebound of the earth's crust and those due to eustatism. Although it is customary to distinguish between eustatic (world-wide) and isostatic (local) changes, the two should not and cannot be sharply separated in that any depression or elevation of the earth's crust close to the coast in some way alters the shape of the ocean basins and thus affects sea-level on a world-wide scale. The effect may be small but this is, in fact, a form of tectono-eustasy.

Since the early days of study of glacio-isostasy, there has been discussion as to whether or not there is a peripheral bulge beyond the ice margin (Daly, 1934): is the depression of the crust beneath the ice to some degree balanced by an up-bulge beyond the ice margin and does this bulge sink when the glaciated area begins to rebound with the dissipation of ice? Geophysicists seem to agree that it should have existed but so far there is little conclusive evidence that such a bulge was formed (Flint, 1957; McGinnis, 1968). The topic is, however, still open to debate (Newman and March, 1968).

More recently, some attention has been given to the concept of hydro-isostasy; the idea that, as eustatic sea-level rose, the weight of water covering the continental shelves caused them to sink (Bloom, 1963, 1965, 1967; Mörner, 1969; Higgins, 1969). This concept and that of a peripheral bulge are mentioned at this point because they complicate the construction of curves of eustatic changes of sea-level.

Such changes have to be considered in this study because the world -wide changes affected the Bay of Fundy, determining the depth of water in it and also its size and shape, all of which influenced the tidal range. As has been shown earlier, the tidal situation must be known in order to interpret accurately morphological indicators of sea-level changes.

With respect to eustatic changes, there are really two things to be established as far as this study is concerned: 1) the maximum fall of sea-level during the last glacial period; and 2) the nature of the rise since that time. This discussion concentrates on the last glacial stage (Wisconsin) and the time since then, because features formed during inter-glacials, which might now be above sea-level, were probably destroyed or modified beyond recognition during the last glacial stage when ice is known to have covered the whole area. Evidence of lower sea-levels during pre -Wisconsin glacials in the form of submarine terraces may exist, but research into this topic requires sophisticated equipment and techniques which were not available to the writer.

Valentin (1954) in his monograph, *Die Küsten der Erde*, includes a useful table listing the estimates for sea-level lowering during the Pleistocene by workers from 1842 onwards. Some estimates made since 1954 are listed in Table 1. Although the estimates vary, most of them lie between about 300 and 450 feet. Inspection of hydrographic charts of the Bay of Fundy and the Gulf of Maine show that, even with the minimum figure quoted in Table 1 - the 295 feet of Kuenen - much of the Bay of Fundy and the Gulf of Maine would have been above sea-level.

Such a statement, of course, ignores the complicating factors that the area was covered by ice at the time, that the earth's crust was depressed by the weight of the ice and that the submarine topography today is different, at least in detail, from what it was during the last glaciation. However, as the Bay of Fundy was, theoretically at least, above sea-level at the last glacial maximum, eustatic rise of sea-level since that time has to be considered when discussing changes of sea-level in the Bay - particularly with respect to changes of depth, size, and shape caused by such changes.

There is as much disagreement about eustatic rise of sea-level since the last glacial maximum as there is about the maximum extent of sea-level lowering. A major point of disagreement is whether or not eustatic sea -level has been higher than it is at present. Dietz and Menard (1951, p. 2015) stated that "there is good evidence that sea-level has lowered

5 - 15 feet in 5,000 years (0.1 - 0.3 foot per century) since the so-called Climatic Optimum", and Bradley (1953, p. 543) claimed that inter-tidal tree stumps found at Robinhood, Maine, "suggest a marked eustatic lowering of sea level immediately after the Climatic Optimum". In addition, Valentin (1954) showed on his graph of eustatic changes a post-glacial high of 6 meters (20 feet) (Fig. 50). Fairbridge (1961, 1962) is probably the leading exponent of the idea that sea-level has been above its present level during the past 6000 years. He believed that it reached its present level about 6000 years ago and has since fluctuated 2 to 3 meters (6 to 9 feet) above and below this level (Figs. 51 and 52). This idea received support in principl from Schofield (1962, 1964) and from Gill (1961), although the former suggested some changes in detail (Schofield, 1964) (Fig. 53).

TABLE 1

SOME ESTIMATES FOR SEA-LEVEL LOWERING DURING THE PLEISTOCENE

Author and Date of Publication	Amount of Lowering (in Feet) and Time at which it Occurred
Kuenen (1954)	383 feet, maximum: 295 feet, the last ice age.
Brannon Jr. *et al.*(1957)	450 feet, last glacial advance (Late Wisconsin.
McFarlan (1961)	450 feet.
Fairbridge (1961)	328 feet, Late Wisconsin.
Curray (1961)	394 feet, 20,000 years ago.
Donn, Farrand and Ewing (1962)	346 feet or 405 feet, Classical Wisconsin.
Cotton (1962)	450 feet, early part of the Wisconsin.
Emery and Garrison (1967)	403 feet, 19,000 years BP
Shepard and Curray (1967)	430+ feet, 50,000 years BP; 380+ feet, 20,000 years BP.
Milliman and Emery (1968)	426 feet, 16,000 years BP.
Guilcher (1969)	328 feet and probably 592 feet, Middle and/or Late Pleistocene.

Another school of thought considers that sea-level has not risen higher than at present, but that it reached its present level some thousands of years ago and has not changed appreciably since then. Godwin, Suggate and Willis (1958) claimed that the sea reached its present level 5,500 years ago and has since remained relatively steady. McFarlan Jr. (1961) took much the same position (Fig. 54), as did Curray (1961) except that he believed the present sea-level was reached 3000 years ago (Fig. 55). Later he modified this position somewhat and stated that "it is sufficient to consider the past 7,000 years as a period of slow rise of sea level followed by relative stability " (Curray, 1965, p. 725).

The other possibility is that sea-level has continued to rise throughout post-glacial time and has never been higher than it is now. The leading proponent of this point of view is Shepard. Shepard and Suess (1956) found no evidence that sea-level was higher during the "Climatic Optimum" than at present; later, on the basis of radiocarbon dates from several parts of the world, Shepard (1960) calculated that sea-level was at -40 feet at 8000 BP and that there has been a slow rise since, although there may have been periods of standstill. This was modified later (Shepard, 1961) to a height of -20 feet at about 6000 BP followed by a slow rise with the sea some six feet below the present during most of the time from 3500 to 2500 BP. Later he showed sea-level reaching a minimum of about -370 feet, 19,000 years ago and suggested that it has risen steadily during the past 15,000 years but at a decreasing rate over the past 6,000 years (Shepard, 1963b) (Figs. 56 and 57). Further examination of the evidence for the last 7000 years (Shepard and Curray, 1967) has led to a suggestion that the sea reached its present level about 2000 years ago, but that it has never been higher than it is today (Fig. 58).

Shepard is by no means the only worker favouring the continuous rise of sea-level theory. Others include Jelgersma (1961, 1966) (Fig. 59) and Šegota (1968) (Fig. 60). It should be noted that, in addition to these studies containing data from many parts of the world, curves have been drawn for sea-level change along the east coast of North America, several of which are believed to support the idea of a continuous eustatic rise over the past 6000 years. These include studies along the Connecticut coast (Bloom and Stuiver, 1963); along the south coast of Florida (Scholl, 1964; Scholl and Stuiver, 1967; Scholl, Craighead, and Stuiver, 1969); and of the coast of Bermuda, southern Florida, North Carolina and Louisana (Redfield, 1967).

So far in this discussion only the two extremes have been considered; that is, the amount of lowering during the last glacial maximum and the situation over the past 6000 years. Of equal importance to the overall study of the Bay of Fundy is the rate of rise of sea-level during the intermediate period. With the development of radiocarbon dating and the increasing availability of it as a research technique, the number of curves of eustatic sea-level changes has proliferated.

As can be seen from the foregoing discussion and from Figures 50 - 60, there is disagreement about the amount of change and the rate of change. In North America at least there are two main schools of thought, one led by Shepard who favours a continuous increase in sea-level since the last glacial maximum perhaps reaching the present level 2000 years ago and the other by Fairbridge who prefers the idea that sea-level reached its present level about 6000 years ago and has fluctuated above and below it since that time. The argument seems to revolve around whose piece of coast is stable.

One school, finding a terrace at say 8 feet above sea-level, will find evidence to show that the coast is unstable and that therefore the terrace does not indicate a eustatic change. The other school finds evidence that the coast is stable and that the terrace, therefore, represents a eustatic rise of eight feet. For example, Scholl (1964) reasoned that geologic and geomorphic evidence points to the fact that Florida is tectonically stable. Therefore, "the +10 ft. Silver Bluff shoreline recognized along the eastern coast of the United States, and its equivalent mapped elsewhere in the world" (Scholl, 1964, p. 347) must be due to eustatic change of sea-level. But eustatic sea-level has not been at a height of +10 feet during the Recent Epoch, therefore the terrace must be of Sangamon (last inter-glacial) or mid-Wisconsin age.

Conversely, Coleman and Smith (1964), working in central coastal Louisiana, found past marsh surfaces now buried at depths ranging from 4 to 40 feet. On the basis of evidence for astandstill in sea-level during the past 2000 - 5000 years, they reasoned that the coast is subsiding at a rate of 0.24 feet per century.

Theoretically, eustatic changes should be studied in stable regions and on this basis most of the east coast of North America can be eliminated. The area that was covered by ice during the last glacial maximum is probably still undergoing isostatic readjustment. This includes an area extending at least as far south as the latitude of New York City. Recently Fairbridge and Newman (1968) found evidence that the New York area is undergoing crustal subsidence. Further south, evidence for crustal subsidence and warping has been found in New Jersey (Stuiver and Daddario, 1963); along the eastern shore of Virginia (Newman and Rusnak, 1965); at the

Previous Work

The emergence of the Fundy coast has long been recognized. Gesner (1861) observed that marls containing relics of shells and sea-weeds which inhabit the present shores are found above sea-level at several localities on the New Brunswick coast between the United States' border and Emerson Creek. The reports of Matthew, Chalmers and Bailey, published during the latter part of the nineteenth century and the early part of the twentieth century made frequent reference to coastal emergence.

Matthew (1875) concluded that, following glaciation, the water surface in southern New Brunswick was 700 feet higher than at present. This figure, now known to be excessive, resulted from misinterpretation of glaciofluvial deposits.

Chalmers (1890, 1893, 1895) listed many features now above sea-level which he believed to be of marine origin and late-glacial or post-glacial in date. In 1890 (pp. 63N - 64N) he published a list of "marine" terraces extending along the New Brunswick coast from the Magaguadavic River to the Petitcodiac River and ranging in height from 75 feet to 230 feet. He concluded that during late-glacial and post-glacial time the land subsided "about 220 feet below its present level relative to the present high tides of the Bay of Fundy" (Chalmers 1890, p. 10N). In 1893, after studying fossiliferous deposits in the Saint John area, he reaffirmed his belief in a subsidence of 220 feet; and in 1895 (pp. 23M - 24M) he published a table showing, among other things, the elevation above mean-tide level of the highest Pleistocene or post-glacial shoreline, at localities in New Brunswick and Nova Scotia, the highest figure being 251.95 feet. There is, of course, no justification for implying such a high degree of accuracy.

the Saint John Valley to a height of 563 feet at Grand Falls. The same height was given by Lougee (1953) and is far in excess of anything estimated by previous workers with the exception of Matthew (1875) in the nineteenth century. To the writer's knowledge there is no evidence that the sea penetrated as far as Grand Falls, although Lee (personal communication) thought there is evidence of marine conditions at Fredericton. It seems likely that the features which Kiewiet de Jonge and Lougee took to be of marine origin were formed in some other way.

On the Nova Scotia side of the bay, work was renewed by MacNeill (1951a), Purdy (1957) and Swayne (1952). Each listed remnants of terraces up to 200 feet above sea-level, but there is no proof that they are of marine origin. Hickox (1958) could find no evidence to support a figure greater than 65 feet for the relative lowering of sea-level along the south shore of the Bay in post-glacial time.

As far as can be judged from reading the literature cited, no corrections were made for eustatic sea-level changes. Consequently, assuming that all the observations are correct, what is recorded in each case is the "marine limit", although this is not always clearly stated. *The Glacial Map of Canada* (Prest, Grant, and Rampton, 1968), a useful summary of what is known about the marine limit around the Bay of Fundy, shows eleven localities at which the approximate marine limit is known, the highest being 230 feet at Lancaster, New Brunswick.

Isobases

Several small-scale maps which attempt to show isobases in the Fundy area have been drawn and all are subject to at least one of the following criticisms: 1) DeGeer (1892, p. 457) defined an isobase as a line of equal deformation. More specifically, Andrews (1970, p. 7) defined an isobase as "a line joining points of equal post-glacial emergence (coastal sites) or equal post-glacial uplift (glacial lake shorelines) where these elevations are the result of post-glacial emergence or uplift operating over the same length of time". In the Fundy area, with very few exceptions, there is no indication of the time element involved. There is, then, little or no evidence that the various uplifted marine features located on the isolines were formed at the same time; 2) the isolines have been drawn using very few control points; 3) some of the control points are of doubtful marine origin; and 4) no allowance is made for eustatic changes.

The first map showing isobases in the Fundy area was drawn by DeGeer in 1892. Attempting to show the whole of eastern North America, he was limited to six control points and drew isobases at 200-foot intervals (Fig. 61).

The map drawn by Fairchild (1918) covered much the same area but in this case isobases were drawn at 100-foot intervals (Fig. 62). The general trend of the isobases on the two maps is the same - roughly parallel to the Fundy trough - but the actual position of the lines varies from one map to the other; for example, the zero isobase on DeGeer's map follows the line of the Annapolis Valley and passes through Truro, whereas the hypothetical zero isobase on Fairchild's map is located, in part, off the south-east coast of Nova Scotia.

Support for DeGeer's map was forthcoming from Daly (1921) who stated that the zero isobase cuts the Nova Scotia coast somewhere between

Yarmouth and Digby and that his observations "confirm the essential accuracy of DeGeer's map " (Daly 1921, p. 388).

In contrast, Goldthwait (1924) produced a map which is closer to that drawn by Fairchild than to that drawn by DeGeer. The zero isobase cuts the south coast of Nova Scotia near Yarmouth and then runs close to the south-east coast. On his map (Fig. 63) Goldthwait showed only five control points on the Fundy coast, one being a height of 60 feet at the apex of Cobequid Bay. This is the height of a feature that he took to be a delta plain representing the limit of post-glacial marine emergence.

Flint (1940) constructed a map somewhat different from those previously published. Showing the Maritime Provinces and Newfoundland, it is entitled *Map showing tentative isobases on 3 former sea-level surfaces suggested by field data*. This is probably the first recognition, for this area at least, of the fact that uplifted marine features which are now at the same height were not necessarily formed at the same time. Tentative isobases were drawn at 50-foot intervals (Fig. 64). There apparently are only four control points for the Fundy area and, like Goldthwait (1924), Flint showed the zero isobase passing east of Truro.

The maps drawn by Lougee (1953) are entirely different. He believed in hinge lines but, although his maps seem somewhat strange, he was aware that features located at the same height are not necessarily synchronous. A simplified version of the conditions that existed during the supposed "DeGeer stability" and isobases on the "DeGeer water plane" are shown in Figure 65. As with Goldthwait (1924) and Flint (1940) the zero isobase is shown passing east of Truro.

Farrand and Gajda (1962, 1965) showed the 100-foot isobase passing through Truro. There were only two control points for the Fundy coast and no claim to accuracy was made, in that both the 100-foot and 200-foot isobases were shown in broken lines and labelled "assumed" (Fig. 66). King (1965) used a contour interval of 50 meters and included most of the Bay of Fundy between 0 and 50, the zero isobase passing well to the east of Truro (Fig. 67). Andrews' (1970, fig. 5-5) isoline map of the marine limit over north-eastern North America shows a 50-meter isoline crossing the bay in a north-south direction.

The most detailed maps published to date are those by Borns (1966) and Swift and Borns (1967b). In the first case, the isobases are shown in feet at 25-foot intervals and the datum is spring high-tide (Fig. 68). On this map the zero isobase is shown passing well to the west of Truro. In the second case, isobases are shown in meters at 10-meter intervals and the datum is mean sea-level (Fig. 69). The trend of the isobases is the same on the two maps but, as the data are different, their positions are changed slightly; for instance, the zero isobase is a few miles further west on the first map than on the second.

Although two reasonably large-scale isobase maps have been constructed for part of the Bay of Fundy, the drawing of further isobase maps for the whole Bay would be premature. Isobases should be drawn only when there are many control points which fulfil the following conditions: 1) physiographic features used as control points should be clearly of marine origin; 2) if they are joined by isobases, the age of formation of the physiographic features should be known; 3) the relationship to sea-level at the time of their formation should be known; and 4) corrections for eustatic changes should have been made.

At present there are few such control points, but given time and careful observation there should be no great difficulty in fulfilling the first condition. The second presents a tangible problem - that of finding some datable material, dating it and relating the date to the physiographic feature. Unfortunately, the most likely material, shells, has two disadvantages: 1) contamination is easy, giving faulty radiocarbon dates; and 2) they fix sea-level very imprecisely, giving a minimum value only. Fulfilling the third condition is difficult for reasons outlined earlier and, at the time of writing, eustatic changes are a topic of hot debate. The problem may be so great that true isobase maps for the Fundy area will not be drawn and we shall have to be staisfied with maps showing isolines for the marine limit. However, if such maps are drawn, there should be a clear statement of what they show.

Description and Analysis of Evidence

In order to show that emergence has occurred, a worker looks for things of marine origin that are now above sea-level. They include fossil: wave-formed cliffs; wave-cut notches; wave-cut platforms; caves; stacks; deltas; beaches and other depositional forms such as bars, spits, and tombolos. Sediments which have no definite physiographic expression but which can be shown to be of marine origin and are now above sea-level are also good indicators of a previously higher sea-level.

Each physiographic form should be carefully studied and its most likely mode of origin determined. This was not possible in all cases, but enough sites were carefully studied for the writer to have serious doubts about the marine origin of some of the "marine features" listed in the earlier literature. It is hard to avoid the conclusion that once something is said to be of marine origin it will continue to be taken for granted for many years afterwards that this is indeed the case. In the early literature there was a tendency to label any steep slope located inland but close to the sea "an old cliff", without consideration being given to other possible modes of origin. Likewise, any piece of relatively flat land became a "raised beach" or "raised shoreline". Another weakness of much early work was the failure to appreciate the height relationship between the various physiographic features and sea-level. The fact that a gravel ridge (beach) is found at a height of 50 feet above present mean sea-level and that the base of a steep slope (sea cliff) is 50 feet above present mean sea-level along some other part of the coast does not indicate that there has been the same relative change of sea-level in both places.

The most definite indicator that a sediment is of marine origin is the presence of shells in it but, even in this situation error is possible. Birds pick up shells and drop them; thus the occasional loose shell sitting on top of an outcrop has to be regarded with suspicion. Another possibility that has to be considered is that shells and sediment were deposited below sea-level during an inter-glacial or inter-stadial period and both were subsequently scraped from the sea floor by a renewed ice advance to be deposited at a point now above sea-level. For example, Grant believed that shells found in stony clay at an altitude of 15-35 feet at Cape St. Mary, Nova Scotia were emplaced in this way, the ice moving onto the coast from the west or north-west.[1] However, if the shells are abundant, are embedded in the sediment, and occur in layers, these problems do not arise. But at all times it has to be borne in mind that the height obtained for a

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[1]GSC - 695.> 38,000 (Lowdon and Blake, Jr., 1968).

shell layer gives only the minimum change of relative sea-level because shells thrown up to the high-water limit tend not to be preserved.

The absence of shells in a deposit does not prove that it is of non -marine origin. Shells are easily broken up if thrown onto a beach and, once lifted above sea-level, they are easily removed by percolating water. In this connection it is worth noting that, as far as is known to the writer, all the shells found around the Fundy coast were in silt or clay: none have been found in sands.

It is not possible in a reconnaissance study of this nature to describe every mile of the coast and to examine it for evidence of emergence. Instead, sample localities are taken and the evidence for emergence is described and evaluated with respect to reliability.

On Deer Island (Fig. 70) two sites are worth considering: one of them is approximately one half-mile north-west of the settlement of Cummings Cove and the other about one half mile south of Hersonville. In the first case, an extensive exposure of sand and gravel layers is underlain by a tenacious khaki-coloured clay with occasional included pebbles. The gravel pit in which the sediments are exposed is located on the east side, the landward side, of a north-south trending ridge. The sand and gravel layers which are exposed at an altitude of between 100 feet and 120 feet dip steeply eastward, away from the nearest sea water, but towards a lowland which would be occupied by the sea if its level were 100 feet higher than at present. It is tempting to conclude that the sediments exposed represent a marine clay covered by a beach deposit. However, since there is nothing to substantiate the idea, it is thought more likely that the clay is a till and that the sand and gravel were deposited by melt water as ice retreated westward or north-westward after having covered the island.

The second site is quite different in nature. It is located in the northern, seaward, side of a hill known as Devils Hand. The physiographic feature in question is a steep slope with its base at an altitude of between 80 feet and 100 feet. Although it might be interpreted as an old sea cliff, there are obstacles to this interpretation, the main one being that, despite the resistant nature of the rock into which the slope is cut, there is no accumulation of boulders at its base, an almost universal arrangement in the case of contemporary sea cliffs cut into resistant rock around the Bay of Fundy.

Despite a search for shells above sea-level, the writer was unable to find any on Deer Island even though the inhabitants of the island claimed that shells have been found at two localities: Calders Head on the south -west side of the island and at Dadd Good Mountain on the east side.

No clear-cut evidence for emergence was found on Campobello Island and, as was mentioned in Chapter 1, Grand Manan Island was not visited. Therefore, the writer cannot cite any clear evidence for emergence of the three islands - Deer Island, Campobello, and Grand Manan - during late- and post-glacial time.

On the New Brunswick mainland there is much more concrete evidence for emergence. Gadd (1969) located shells near Benson Corner. They have a radiocarbon date of 12,300± 160 (GSC - 886) and were found at an altitude of 85 feet, indicating the minimum figure for emergence. Nearby, on the east side of the St. Croix River, at Sand Point, about five miles north of St. Andrews, there is another shell-bearing deposit. This site was located by the writer in 1965 and shells from it collected in 1966 were identified by Dr. Wagner of the Atlantic Oceanographic Laboratory, Bedford Institute (Appendix 1, sample 1). Unfortunately, not enough shells were collected for radiocarbon dating, but during the summer of 1967 Gadd visited the site and collected a large enough sample. The date obtained was 12,300± 160 (GSC - 795), exactly the same as that for shells at Benson Corner and very close to that for shells found at another locality in the St. Croix Valley (GSC - 1067: 12,600± 170). This indicates that Oak Bay, the Lower St. Croix River and Passamaquoddy Bay were ice free by about 12,300 years ago.

At Sand Point excavation for gravel was taking place when the site was visited by the writer. There was, therefore, a good clear section (Fig. 71). At the base of it there are layers of glaciofluvial gravel dipping to the south or south-west. These were presumably deposited by water issuing from an ice lobe which retreated northward along the line of the Oak Bay - Lower St. Croix Valley. Above the glaciofluvial deposits a grey silt clay which contains the shells, truncates the glaciofluvial deposits and above that is a brown clay containing some shell casts. At the top of the section there are almost horizontal boulder beds which probably represent offlap conditions as relative sea-level fell.

The height of the shell-bearing layer was measured using a tape and clinometer and it was found to be 46 feet above high-water level as indicated by the debris line. It can, therefore, be concluded that about 12,300 years ago relative sea-level was at least 46 feet higher than present high-water level. If this were the case, it is reasonable to expect to find further evidence of marine deposits in the Oak Bay - Lower St. Croix area. There are numerous exposures in gravel pits between Benson Corner and St. Stephen, but the stratigraphy in most cases is complex with deposits of till, ice-contact stratified drift, some of which may have been disturbed by marine action, and some deposits which are probably truly marine in origin. Gadd (1970b) mapped the Quaternary geology of the St. Stephen and St. George map areas on a scale of 1:50,000 and showed "marine basin deposits" at heights up to 150 feet; also he found shell fragments in clays approximately 100 feet above sea-level at Sand Point (personal communication). A minimum figure of 150 feet can therefore be taken for the marine limit in the St. Stephen - St. Andrews region.

Chamcook Mountain, located approximately 2 miles south-east of Sand Cove, has an extremely steep western side (Fig. 72) the base of which is located at an altitude of between 150 feet and 200 feet. The steep face might, therefore, be a sea cliff formed at or about the time that marine deposits were laid down at Sand Point and at other localities on the east side of the Lower St. Croix River. On closer inspection it was found that at the base of the cliff there is an abundance of very large angular boulders (Fig. 73) which have clearly fallen from above. The general appearance of the base of the cliff is much like that of the base of the present sea cliff at Owl Head (Fig. 74) on the west coast of Chignecto Bay. There are, however, two main differences: first, the boulders at the base of Chamcook Mountain are angular compared with those at Owl Head; and second, there is no transition from boulders to pebbles to sand west from the base of Chamcook Mountain towards the sea whereas there is this transition at Owl Head, although sand is certainly not abundant. The conclusion drawn is that the west side of Chamcook Mountain does not owe its steepness primarily to coastal erosion. A more likely explanation is that the slope was steepened by an ice lobe that moved down the Oak Bay - Lower St. Croix Valley, but it is possible that for a short time following deglaciation when sea-level was at the marine limit the west side of Chamcook Mountain formed the coast.

Further east in the area covered by the 1:50,000 St. George West (21G/2W) topographic map (Fig. 3), there are several other examples of steep slopes cut into resistant rock which might be taken as old cliffs. The most noticeable are: 1) west of Bocabec River near the settlement of Bocabec; 2) on the west side of Digdeguash Harbour; 3) the north-west slope of Rocky Hill, north of Blacks Harbour; and 4) on the south-east side of Hawkins Hill. They all fit the criteria of being steep slopes, close to the sea with the base of the slope being above present sea-level— the sort of thing that was automatically labelled an "uplifted cliff" in earlier literature. Supporting evidence for this identification is minimal; for example, little indication of any beach material exists at the base of any of the "cliffs" although in the case of Rocky Hill a few large boulders can be found. At Hawkins Hill there is a flat area to the south-east of the "cliff". This area, which is less than 50 feet above sea -level, could be interpreted as an uplifted wave-cut platform.

On his map of the Quaternary geology of the St. George map area, Gadd (1970b) plotted "marine basin deposits" up to a height of 200 feet. If this is taken as a minimum for the marine limit, the four slopes mentioned were at some time during the Quaternary washed by the sea, which might have contributed to their steepness.

In the area covered by the 1:50,000 St. George East (21G/2E) topographic map evidence of emergence is more definite, taking the form of a large outwash fan/delta complex. The topography is distinctive. To the north in the area of Jack Lee Mountain the land is hilly, rising to heights of over 800 feet in places. It decreases in height in a general southerly direction and there is a marked southward projection of the 250-foot contour line, the path of which gives a good indication of the general shape but not the exact boundary of the outwash fan/delta complex (Fig. 75). To the north and east of Utopia Army Camp are several enclosed depressions, some of which are occupied by lakes - for example Jerry Pond and Knight Pond - and some of which are dry. Another feature of the topography is a great extent of flat or very gently sloping land between the 300-foot and 200-foot contour lines. Finally, several abandoned scarps, none of which is very high, are very prominent elements of the landscape as a result of the general flatness just mentioned. They show up very clearly on air photographs of the area and in one case, south of Highway 1 at Pennfield Ridge, a low ridge extends in an east-west direction for about a mile at the top of one of the scarps.

By inspection of gravel pits which are abundant in this area, it can be shown that the Pennfield Plain (roughly the area between the 200-foot and 300-foot contour lines) is underlain by surficial deposits. The total thickness of surficial material varies from less than 25 feet in the north near the Messenett Stream to over 200 feet south of Pennfield Ridge (Gagné, 1969). It consists of glacial till, marine clay and glacio-fluvial deposits, the latter being composed in the main of sand and gravel layers, usually with a southerly component of dip. In some exposures these layers are overlain by thin, nearly horizontal gravel beds which are interpreted as the result of the reworking of the glaciofluvial material by the sea. This arrangement is found at sites 1, 2, and 3 (Fig. 75) at altitudes greater than 200 feet. Therefore, the marine limit in the Pennfield Plain area is greater than 200 feet and, as Gadd (1970b) mapped "wave modified outwash fan or delta" at altitudes over 250 feet, it is probably greater than 250 feet, though the writer has personally seen nothing in the field to prove this.

The stratigraphic arrangement is probably best illustrated at site 4 (Fig. 75), where the section consists of glaciofluvial materials below with a southerly component of dip, overlain by silt and then by a thin band of red clay (Fig. 76). Above this, bands of sand and fine gravel make up the western end of the east-west ridge mentioned earlier. They are nearly horizontal or even have a slight northerly component of dip. The ridge is, therefore, believed to be a storm beach thrown up when the sea was approximately 200 feet higher than at present.

The scarps marked on Figure 75 are probably part depositional and part erosional in origin. Although they may represent the leading edge of the foreset beds as the delta extended southward, it is very likely that they were steepened by wave erosion when the sea retreated to the south as emergence took place.

Shells have been found a mile south-west of Pennfield Corner in clay "that probably forms bottom-set beds of the Pennfield glaciofluvial delta" (Gadd, 1969, p. 195). They were dated 13,000± 240 years BP (GSC - 882), indicating that ice had withdrawn from this area by that time. The depressions near the Utopia Army Camp are probably due to the melting of buried ice remnants after the withdrawal of the main ice mass.

The exact relationship between the date of deposition of the shells and the date of highest relative sea-level in the area is not known. Since the shells were found at an altitude of 130 feet, it is safe to say that about 13,000 years ago relative sea-level was at least 130 feet higher than it is at present and, based on the interpretation of the deposits underlying Pennfield Plain, that at sometime since the area was deglaciated it has been at least 200 feet and probably 250 feet higher. The date obtained for the shells gives a minimum for the deposition of glacial material to the north.[1]

In the area covered by the 1:50,000 Musquash West (21G/1W) topographic map, three sites were investigated. On the south side of Little Lepreau Basin there is a small sand and gravel pit. The sand and gravel layers dip steeply (up to 32^{o}) to the north, that is towards Little Lepreau Basin, and are plastered against a bedrock face to the south. The steeply dipping layers are truncated at the top by an almost horizontal surface at a height of approximately 50 feet (taken from the 1:50,000 map). The sand and gravel layers are thought to be of glaciofluvial origin and the 50-foot terrace surface was formed when water level in Little Lepreau Basin was 50 feet higher than at present. It was not possible to determine exactly when this was, but it is reasonably certain that it was less than 13,000 years ago, the date when Pennfield Plain was free of ice.

A somewhat similar arrangement exists on the south side of a small stream flowing into the west end of Dipper Harbour where a small pit has been cut into gravel, the layering of which is nearly horizontal when viewed from the north, that is from the direction of the stream. The top

1A date of 16,500± 370 (GSC - 1063) for lacustrine organic material (gyttja) obtained from a kettle lake in unmodified glacial features may indicate that the ice margin retreated from the present coast long before the limit of marine submergence was reached (Gadd, 1970a).

of the gravel layers coincides at a height of 49 feet (measured by altimeter with the debris line as datum) with a well-marked terrace which continues eastward, most of the settlement of Dipper Harbour being built on it. It is interpreted as an indicator that the sea-level was once 50 feet higher than at present in Dipper Harbour. The height of land between Little Lepreau Basin and Dipper Harbour is over 50 feet, but less than 100 feet; so in all probability the two water bodies were connected in the past and the terraces just described may be genetically related.

The third site is on the east side of Musquash River, 750 feet north of Highway 1. Here an exposure of fine-grained gravel and sand about 30 feet high is cut into a terrace, the back of which is 90 feet above the debris line on Musquash Marsh. Dipping towards the Musquash River, the sand and gravel layers are topped by a layer of red clay. This clay is thought to have been deposited and the terrace formed when the Musquash River, and by implication sea-level, was as much as 90 feet higher than at present.

In the Lorneville area there is some evidence of emergence. A small stream flows north-eastward into Lorneville Cove and between it and the coast a north-east/south-west trending ridge reaches a height of 200 feet. Sand and gravel deposits, found on the north-west side of the ridge up to heights of 200 feet, are layered, the dip of the upper layers being in a northerly direction, towards the stream. In the writer's opinion these are marine (beach) deposits laid down when the sea in this area was at least 200 feet higher than it is at present.

The surficial geology and Quaternary history of the area on either side of the mouth of the Saint John River, an area extending from Lorneville

Point to Black Point, is extremely complex and could alone form the topic of a doctoral thesis. The Saint John River enters the sea, following the most unlikely route, at the Reversing Falls (Fig. 77).

It is unlikely that the river followed this route during pre-glacial time and several alternative routes have been suggested. Chalmers (1890) believed that in pre-glacial time the river flowed from Lee Cove south-west to Manawagonish Cove. He also suggested three courses, other than the present one, that the river may have taken during post-glacial time: 1) the pre-glacial route just mentioned; 2) from Drury Cove via Marsh creek to Courtenay Bay; and 3) from South Bay to Manawagonish Cove. The diversion of the river from its former course or courses is almost certainly connected with glacial deposition and changes of sea-level. Likewise, the river has affected the physiographic form assumed by some of the glacial and post -glacial deposits.

Almost the entire coast between Taylor Peninsula and Black Point is developed on surficial material, although there are some bedrock outcrops, for example, at Sheldon Point. The bedrock is mantled by a complex system of deposits of glacial, glaciofluvial and glaciomarine origin, together with recent marine and fluvial deposits. This system extends in a south -west/north-east direction from Taylor Peninsula through Saint John and north-east of the city roughly parallel to Kennebecasis Bay. Although the whole complex may represent the terminal deposits of ice lobes moving down the Saint John and Kennebecasis valleys, it has been shown (Melvin, 1966) that the local underlying topography played a large part in the distribution of the various deposits in the Ben Lomond area north-east of Saint John.

It is not the writer's aim to describe the distribution of the surficial deposits of the area, but an attempt has to be made to relate marine deposits indicating higher sea-levels to the glacial and glacio-fluvial deposits in order that a chronology may be arrived at. The emphasis is, however, on evidence of the amount and date of marine emergence.

As has been pointed out, the best indicator of emergence is the finding of shells in positions now above sea-level. They are particularly useful in that, if found in sufficient quantity, a radiocarbon date can be obtained. Fortunately, there are several fossil localities in the area being discussed. Between Taylor Peninsula and Sheldon Point the coast, which trends in an east-west direction, is backed by cliffs cut into surficial material; an exception is a stretch of about half a mile east of Taylor Peninsula where there are no cliffs.

At their western extremity the cliffs are low-lying, being no more than 10 feet high, but they increase steadily in height until they are nearly 100 feet high at a point a quarter-mile to the east (Fig. 78). In the west, the cliffs are cut in a red clay which is tenacious when wet but dries to a hard shale. The low cliffs are under more or less constant attack by the sea and slumped masses are common. Where the cliffs are highest the clay is not so pure and includes large numbers of small embedded pebbles. In addition, on the beach are some very large boulders which might have been washed from the clay (Fig. 10).

Shells are abundant in the clays. One sample was collected at site 1 (Fig. 78) from a slumped mass which included layers containing shells, some of which were whole and therefore easily identifiable (Appendix 1, sample 2). The cliff height at this locality is only 20 feet so when *in situ* the shell layer could not have been more than this amount above sea-level. At site 2, where the cliffs are much higher, shell fragments were collected from the clay at heights as great as 59 feet above the cliff base (height measured by tape and clinometer). Unfortunately, the fragments were small and no identifications were possible.

Shells collected by Lee from the general vicinity of site 1 have been dated at 13,325± 500 [I(GSC) - 7]. Later, shells collected by the writer from site 3 were run as a check on this date, the result being almost the same: 13,200± 200 (GSC - 965). The shells collected by the writer came from red clay, just below high-tide level, and it was noticeable that the shells are concentrated within thin layers of black clay, only casts being preserved in the red clay. Identifications are given in Appendix 1, sample 3. Lee stated that the clay from which his shells were taken lies beneath and intertongues with delta-outwash gravels and that the shells were deposited during or shortly after the retreat of the glacial margin from the area. However, Gadd (1970a, p. 170) stated that the feature which Lee took to be a glaciomarine delta "is a submerged end moraine modified by marine deposition and wave action". The data for the shells gives the date at which they were deposited and ice may have disappeared from the area at this time or shortly before.

The peninsula lying between Manawagonish Creek, Sand Cove, and Sheldon Point is underlain by surficial material, the general arrangement of which can be seen at sites 3 and 4. At site 4, running in a north-south direction, is a gravel excavation cut some 30 feet below the ground surface. At the bottom of the exposure, gravel layers containing occasional large boulders (Fig. 79) dip steeply to the north. At the top of the exposure are nearly horizontal gravel layers. At site 3, close to the present coast, there are again well-layered gravel deposits but in this case the dip is towards the south. However, at the top of the exposure is an almost horizontal layer of very coarse gravel and boulders (some up to 2 feet in diameter). At one point, at the extreme southern end of the exposure, red fossiliferous clay is pinched out between the steeply-dipping pebble layers below and the nearly horizontal beds above (Fig. 80).

The sequence of events suggested for this area is as follows. The steeply-dipping pebble layers were deposited first, under glaciofluvial conditions. The exact depositional conditions are not obvious, but neither the glaciomarine delta of Lee, nor the submerged end moraine of Gadd seems to explain the opposite direction of dip of the gravel layers on opposite sides of the topographic axis of the peninsula. However, once they were deposited, relative sea-level rose and fossiliferous red clay was deposited around the gravel in the deeper water, while in the shallower water wave action sorted the upper gravel layers and re-deposited them in almost horizontal layers. This interpretation of events is closer to that suggested by Gadd (1970a) than to that implied by Lee. If it is correct it means that at some time after about 13,200 years ago sea-level reached at least 100 feet higher than it is at present, for the horizontal, wave -sorted, gravel layers are located at least as high as 100 feet above present sea-level.

Chalmers (1893) referred to the occurrence of marine shells in boulder clay on the west side of Saint John Harbour between Negrotown Point and Bay Shore. This section of coast was examined by the writer with the purpose of trying to collect shells which could later be identified and radiocarbon dated. The cliffs in this area, in the order of 50 feet high, are cut into a red clay impregnated with boulders, some of which are several feet in diameter. Their enormous size is best seen when they have been washed from the cliff and have accumulated on the beach below (Fig. 81). The lithology of the boulders is exceedingly varied. Shells were found at several localities, but the best site is located approximately a quarter -mile west of the breakwater at Negrotown Point. Here shells were taken from a clay band approximately 6 feet above the cliff base and have since been identified (Appendix 1, samples 4 and 5). The shells were very well preserved, whole valves were collected in some cases. Elsewhere sections are not too clear because of the slumping of the clays and the fact that the top of the cliff is obscured by waste materials, presumably dumped from the railway sidings at the top of the cliffs. However, at one spot about a quarter-mile west of the fossil locality gravel layers, which are some 4 feet to 5 feet thick and almost horizontal when viewed from the south, can be observed at the top of the cliff (Fig. 82).

The writer's interpretation of the stratigraphy described is as follows. The red boulder clay with its included layers of fossiliferous clay must have been deposited below water. As the shells occur in abundance, in layers, and are well-preserved, it is thought that they are *in situ*. Also, as the boulders within the clay cause no depression of bedding planes; that is, there are no "splash marks", it is thought that they were not dropped into the clay from rafted ice blocks. The boulder clay sequence probably represents part of a submerged end moraine, the shell-bearing layers being deposited on the moraine during periods of quiescence. A date for the shells would, therefore, fix the maximum date of retreat of ice from this area, but unfortunately none is available. The horizontal gravel layers on top of the boulder clay were deposited by wave action in shallow water some time after the boulder clay was deposited. Sea-level has, therefore, been at least 50 feet - the altitude of the gravel layers - higher than at present at some time since ice retreated from the area. There may be some genetic connection between the surficial deposits of the Negrotown Point - Bay Shore area and those of the Sheldon Point Peninsula, but in the absence of a detailed map of the surficial geology of this area the writer is not prepared to speculate further.

Inland, roughly bounded by the 150 feet contour line, is a ridge which extends south-west from Quinton Heights to a point north of Taylor Peninsula where it swings westward (Fig. 77). About a mile north-west of Taylor Peninsula there is a break in the ridge at a point where it is crossed by a suggested former course of the Saint John River from South Bay to Manawagonish Cove (Chalmers, 1890; Alcock, 1938). However, the col is over 100 feet above sea-level and, if the river did follow this course, water in South Bay must have been at least 100 feet higher than it is at present, unless the col has been partially filled since water occupied it.

As far as can be told by inspection of exposures in gravel pits, the ridge is composed completely of surficial materials. The best exposures occur in two gravel pits at the west end of the ridge - one on its north side (Fig. 77, site 5) and one on the south side (Fig. 77, site 6). At site 5 excavations have taken place at two levels (Fig. 83). In the lower level of the pit almost all the material exposed is gravel, from fine to coarse. This gravel is well-layered and the dip is steep (up to 30°) and to the north. In the upper level of the pit conditions are more variable. At the bottom of the exposure there are gravel deposits which might be the same as those in the lower level of the pit. However, they are in the main covered by material which has slumped from above. Thus, the exact arrangement is difficult to determine. Above the gravel a horizontally-bedded sand layer reaches a thickness of 4 feet in one place. Although the sand is mainly of uniform grain size, it does include a few boulders. Above the sand is a red clay; irregular in thickness, it reaches 4 feet in some places and contains pebbles, small boulders and occasional shell fragments, none of which was large enough to be identified. At the top of the exposure there are horizontal layers of gravel at least 10 feet thick.

At site 6 on the south side of the ridge the arrangement is somewhat similar, although not so clear. At the bottom of the exposure gravel layers, very similar to those at the bottom of the sequence at site 5, are exposed. They dip steeply - up to 26° (Fig. 84) - but in this case to the south rather than to the north. At the east end of the exposure, red clay can be seen overlying the steeply-dipping gravel layers and it in turn is overlain by more gravel.

The interpretation of the arrangement at sites 5 and 6 is as follows. The steeply-dipping gravel layers at the base of each exposure are of glaciofluvial origin, although the writer is unable to offer a reasonable explanation for the fact that they dip in opposite directions on either side of the ridge. At a later date, sea-level rose and sand was deposited at site 5 and clay at both sites. The horizontal gravel layers at site 5 may represent beach deposits as the sea became shallower.

The shell-carrying clay is at an altitude of between 100 feet and 150 feet at both sites, so it is safe to conclude that sea-level was at least 100 feet above present level at some stage after the retreat of ice from the area.

It is worth noting at this point that the north side of the ridge is scarp-like in form, the base of the scarp coinciding closely with the 100-foot contour line. If at some stage in the past, after the deposition of the materials that now form the ridge, water in South Bay was over 100 feet above its present level, as it would have had to be for the Saint John to flow along the South Bay - Manawagonish Cove route, then this scarp would have formed the coastline.

Further evidence of marine emergence can be found on the east side of Saint John Harbour between Little River and Black Point. At Red Head the cliffs are cut into red clay. The wave-cut platform is underlain by red-brown clay and on it lie many boulders which, although large, are not as big as those at Negrotown Point. The cliffs, however, are similar in form and composition (Fig. 85). From the base of the cliff to a height of about 10 feet the material exposed is red clay containing some large boulders. Probably the boulders on the wave-cut platform have been washed from the clay: the cliff is clearly being actively eroded, for the bottom 15-20 feet is almost vertical although cut in extremely mobile material. Above the "boulder clay" there is about 10 feet of clay which contains occasional small pebbles. Shells are abundant in this clay and are sufficiently well-preserved for identifications to be made (Appendix 1, sample 6). At the top of the exposure are sands and gravels and behind the cliffs is a small sand and gravel pit.

About 2000 feet south of the point of Red Head cliffs are again cut into red clay, but here the clay is nearly all of the "boulder clay" variety, some of the boulders being as much as 5 feet in diameter. However, within the "boulder clay" are thin layers of clay which contain shells. One such layer is located 5 feet above the cliff base and sufficient well -preserved shells were collected from it for identification (Appendix 1, sample 7).

Cliffs, composed of red clay below with sands and gravels above, extend for just over a mile south-east from Red Head, and shell fragments can be found in the clay at several points. Although the cliffs are in the order of 40 or 50 feet high along most of this stretch, they decrease in height to the south-east. The clays are for the most part overlain by sand and gravel and, when any layering of these can be distinguished, it is nearly horizontal. A spring line exists at the junction of the sands and gravels and the clay, with the result that the clay is extremely wet and mobile. Slumped masses are common, but these in turn are rapidly eroded by the sea to leave behind the boulder content which decreases in coarseness from north-west to south-east. The existence of red clay in the cliffs of the Red Head area at a height of 40 feet is taken to indicate at least 40 feet of emergence.

The deposits in the Red Head area bear a remarkable similarity to those found at Negrotown Point and a similar explanation of their formation is suggested; that is, that the "boulder clay" represents a submarine moraine and that shell bands within it represent periods of inactivity of the ice when quiet marine conditions prevailed. Indeed it is possible that the moraine was once continuous across Saint John Harbour and that it was breached when the Saint John River was forced to change from one of its old courses.

Some indications of emergence are found inland. Shell fragments were found at the bottom of a large sand pit located on the south side of

Little River, south-east of Silver Falls (Fig. 77). The fragments could not be identified and there was not enough material for radiocarbon dating. The shell fragments were found at a height of between 100 and 150 feet (height taken from 1:50,000 topographic map). Therefore, at some stage in the past, sea-level has been at least 100 feet higher than it is at present. Two miles south at Midwood, Melvin (1966) collected microfossils at an altitude of about 40 feet. He believed that a glaciomarine delta was deposited in the Little River area and stated that fossiliferous marine clay at the base of the deltaic foreset beds represents the bottomset beds. He mapped deltaic deposits to a height of over 150 feet. Thus, if his interpretation is correct, sea-level was probably at least 150 feet higher at some stage in the past than it is now and certainly more than 40 feet higher than at present.

The shell samples collected by the writer from sites on either side of Saint John Harbour have a wide enough geographic range that they do not suggest any specific temperature. They can be found presently off the coast of New Brunswick or they can be found further north. The dominant species of foram. in most of the samples, *Elphidium incertum clavatum*, points to a shallow inner-shelf or open-bay environment (Wagner, personal communication).

Matthew (1879) reported that Leda Clay[1] is exposed near Lawlor Lake, about 4 miles north of Saint John. The clay contains *Portlandia glacialis*,

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[1] The terms "Leda Clay" and "Saxicava Sand" are used in some of the earlier literature about New Brunswick. The terms were originally applied to the fine-grained marine sediments (Leda Clay) and coarse-grained marine sediments (Saxicava Sand) deposited in the Champlain Sea (Dawson, 1893). By extrapolation the names were applied to clay and sand deposits found in New Brunswick. There is, however, no evidence of any relationship either in form or time of deposition. The terms are, therefore, best not used.

Saxicava rugosa, and *Astarte Banksii*. The writer did not locate the exposure but, as Lawlor Lake is at a height greater than 100 feet above sea-level (taken from 1:50,000 map), it can be concluded that at some time water level in Kennebecasis Bay was at least 100 feet higher than at present. This would be high enough for the water to flow to the open sea *via* the Drury Cove - Marsh Creek - Courtenay Bay route, assuming (of course) that there was at that time no obstruction which has since disappeared. In fact if the water level in Kennebecasis Bay were to rise by only 50 feet, water would enter the sea *via* this route.

The fossiliferous red clay is almost, but not quite confined to the sides of Saint John Harbour. A small exposure was found at the base of cliffs located between the mouth of Mispec River and Mispec Park. The clay contains shell fragments and, as far as is known to the writer, this is the furthest from the mouth of the Bay of Fundy that shells have been found. From about this point north-east along the New Brunswick Fundy coast the geomorphologist has to look for erosional evidence of emergence, and also for depositional evidence, but there are no shells to confirm the marine origin of the deposits.

A quarter-mile north of the mouth of Mispec River there is possible evidence of marine emergence. Here a terrace with its surface at between 100 feet and 150 feet (heights taken from the 1:50,000 map) is backed by a steep slope cut in bedrock. The terrace has been excavated for sand and gravel and these deposits can be seen to dip in an easterly direction towards the nearest water. The terrace may represent a former high sea -level at between 100 and 150 feet.

Although no shells are found north-east of Mispec Park, this is not to say that the red clay disappears also. About midway between the mouths of Black River and Emerson Creek the cliffs, which are in the order of 50 feet high, are composed of bedrock below with red clay above and sand and gravel at the top. Inland from the cliff there is an area of flat land backed by a scarp, the base of which is more than 50 feet but less than 100 feet above sea-level. The flat area is developed on surficial material and the scarp closely follows the trend of the 100-foot contour line. Therefore, the landforms are interpreted as an uplifted sea cliff and a wave-cut platform covered by marine deposits. They indicate that there was a still-stand of the sea at a height greater than 50 feet above present sea-level; or perhaps more correctly that, when it was at this height relative to the land, sea-level was changing only slowly. It is noteworthy that the area to the north of the scarp (old cliff) has been cleared and is cultivated, indicating that the scarp is cut into surficial deposits. At this point it is worth stressing that at no place along the Fundy coast described so far is there a clear example of an uplifted cliff or terrace (wave-cut platform) cut into bedrock.

In the St. Martin's area between Honeycomb Point and Echo Cove further evidence exists of coastal emergence (Fig. 86). North-west of Honeycomb Point an area of land rises gently inland to a scarp which is thought to be an old cliff because: 1) there is a sharp change of slope that would be expected at the cliff/wave-cut platform junction; 2) there is an accumulation of rounded boulders at the base of the scarp; and 3) the scarp runs nearly parallel to the present coastline. The height of the change of slope was measured using an altimeter (with the debris line on the marsh near West Quaco as datum) and found to be 155 feet. Triassic sediments are exposed in the area between Honeycomb Point and Quaco Head and, since sediments of this age exposed around the Bay of Fundy are usually less resistant than rocks of earlier periods, it is possible that the scarp (old cliff) coincides with the junction of Triassic rock and more resistant older rocks. However, if Geological Survey of Canada Map 477A, *Loch Lomond (East Half)*, is correct the wave-cut platform truncates Triassic and much older rocks and the cliff line is developed well inland from the contact zone between the Triassic and the Proterozoic, though not as far inland as the Proterozoic/Lancaster Formation contact.

North of St. Martins at a height of between 100 and 150 feet is a terrace underlain by surficial material which is probably marine deposited, at least in part. The material consists of gravel, some of which is in layers dipping steeply to the south and is thought to be of glaciofluvial origin, and some of which is in nearly horizontal layers thought to be marine-deposited.

North of Macomber Point there is evidence of another marine deposit, a gravel pit in which gravel layers, remarkably uniform in thickness and dip (direction 190^{o}-200^{o}, amount 8^{o}), are exposed. The height of the top gravel layer is 122 feet (measurement by altimeter from the debris line). To the north-east of this site there is a low scarp about 600 feet inland from the present cliff. It was located by air-photo interpretation after fieldwork had been completed and was not inspected, therefore, in the field by the writer. However, it runs parallel to the present cliffline and, as it is developed on non-resistant Triassic rocks and coincides with no lithological boundary, it is possibly of marine origin. Its base is at a height of between 150 and 200 feet.

At a much lower level 1500 feet south of West Quaco further evidence of emergence exists. On the south side of Mosher River there is an area of flat land underlain by sand and gravel which appears to be of marine origin and was probably deposited when the sea was high enough to occupy the area now covered by the Quaco marshes. The top of the gravel surface exposed on the south side of the Mosher River is 34 feet above the debris line and probably represents a still-stand (or at least only a slow change) of sea-level.

Summarizing the St. Martins area, it can be said that sea-level once stood at least 155 feet higher than present high-tide level and that there is some evidence for periods of still-stand at between 100 and 150 feet and at about 34 feet. There is, at present, no way of dating these events.

Scarps and terraces which may be of marine origin are found: at Fownes Head, between Echo Cove and Martin Head; to the west of the mouth of Big Salmon River; at Tufts Point; and at Seely Beach. It is difficult to reach these sections of coast except from the sea, the localities mentioned being picked out on air photographs only after fieldwork had been completed. As the scarps and terraces are not clearly of marine origin, no attempt was made to determine their altitude. However, their existence is mentioned here so that they can be investigated in detail in any further study.

Between Martin Head and Point Wolfe the coast, which is developed almost entirely on resistant Precambrian rocks, is extremely steep. Access from the land is generally very difficult although Martin Head can be reached more easily. Therefore, apart from the one exception of Martin Head, this stretch of coast was not seen in the field by the writer. However, from the point of view of determining sea-level changes this is not too serious an omission because it is unlikely that any record of higher sea -levels would be preserved on a steep coast developed on resistant rocks. This impression was confirmed by inspection of air photographs; no clear scarps or terraces which might have been of marine origin were observed.

Near the mouth of Wolfe River, four-tenths of a mile inland from Point Wolfe, there is a sand and shingle bar adjusted in height to present sea-level conditions. Three-tenths of a mile further inland an uplifted sand and shingle bar extends across the path of Wolfe River. The top of this bar is 115 feet above sea-level (measurement by altimeter with the debris line as datum). In this particular locality, therefore, the highest indication of a past sea-level is 115 feet above the highest indication of present sea-level. Borns and Swift *in* Poole (1966) gave a figure of 135 feet in the general area of Fundy National Park, but the small scale of the map on which it is plotted makes exact location of the site difficult.

In the area between Alma and Cape Enragé there is further evidence of emergence. The Upper Salmon (Alma) River flows swiftly in a steep-sided valley, but near its mouth there is a marked river terrace on the east side of the valley. Close to the river mouth, three-quarters of a mile north-west of Alma, the front of the terrace is at a height of 58 feet above present high-water level. This shows that water in the river has been 58 feet higher in the past and, by implication, that sea-level has been almost as high; levelling along the terrace would give some idea of the river's gradient at the time the terrace was formed. East of the mouth of the river behind the settlement of Alma there are two low scarps; their bases, below 50 feet, are taken as indicators of former sea-levels.

Some four and a half miles to the north-east, at the top of the cliffs at Dennis Beach, what is probably an uplifted beach deposit can be seen. The cliffs at the western end of Dennis Beach and at Red Head are composed of red Triassic rock (Fig. 15). Between these points the bedrock is overlain by sands and gravels with nearly horizontal layering when seen from the south. Since the cliffs are being rapidly eroded by the sea, the sand and gravel have fallen down the cliff face to obscure the bedrock totally in some areas. The sand and gravel layers were probably deposited when sea-level was higher by an amount of 78 feet (measurement by altimeter using the present debris line as datum). Although there is no clear proof of this, the easily-eroded Triassic rocks were probably bevelled by the sea to form a wave-cut platform, a beach later being formed on top.

Further east there are other clear indications of emergence. Behind the marsh at Anderson Hollow a low scarp is cut into gravel deposits which are thought to be the beach deposits of a once higher sea-level. At one locality, 1500 feet west of Little Rocher, where the gravels underlie an expanse of flat land, their surface is 24 feet above the present debris line. Also, on the east side of a small creek draining to the eastern side of Salisbury Bay, there is exposure of beach-deposited material, this time at something over 50 feet above sea-level.

The village of Albert is built on an uplifted deltaic deposit which, since its deposition and uplift, has been dissected by Shepody River and some of its small tributaries. The best exposure of the deltaic deposits is found in the yard of the Department of Public Works. The foreset beds can be seen to dip almost due south at an angle of 26 degrees and above these there is 4 feet of topset beds (Fig. 87). Using an altimeter and the nearest bench mark as datum[1], the writer found the surface of the delta at this one place to be 64 feet above sea-level.

The delta deposit at Albert is the northernmost indicator of emergence found on the Fundy coast: nothing was found around Shepody Bay or Cumberland Basin. Evidence of changes of sea-level within these two water bodies might be obtained by inspection of the valleys of the rivers flowing into them. Changes of base level may be reflected in the development of river terraces. However, because of limited time and the large size of the area covered by the study, this line of investigation was not pursued.

No evidence of emergence on the Nova Scotia side of Chignecto Bay is found, although between Ragged Point and Sand River a series of ridges, in some places nearly parallel to the coast, might on first impression be taken to be marine-formed features. They are difficult to reach and investigate in the field, but it can be determined almost conclusively by inspection of air photographs that they are not of marine origin. Although in some parts they are nearly parallel to the coast, between Cranberry Head and the mouth of Shulie River they form an acute angle with the cliff line and some of them form marked headlands (Fig. 22). The ridges are, then, determined by variations in bedrock lithology; they are truncated by the coast and a wave-cut platform is developed across them. This seems to suggest that sea-level has been at or close to its present level for a long period of time, but most other evidence suggests that sea-level has been changing during the Recent Epoch. The wave-cut platform may, therefore,

[1] All bench marks referred to in this thesis are based on Canadian Geodetic Datum which is based on the value of mean sea-level. Therefore the height of any feature obtained by using a bench mark as datum is its height above mean sea-level.

be exhumed, having been formed at some time in the past when relative sea -level was about the same as it is now.

The north shore of Minas Basin is quite different in that there is abundant and clear evidence of emergence. Swift and Borns (1967b) recognized what they termed the "Five Islands Formation", an outwash deposit which forms a discontinuous terrace between the Cobequid Mountains and Minas Basin (Fig. 88). Within this formation, they distinguished three lithosomes: 1) A glaciolittoral lithosome which was deposited in a wave-agitated environment and can now be seen in the Advocate Harbour area; 2) A glacio-deltaic lithosome which fills river valleys between Cape Spencer and Five Islands. The deltas have a well-defined topset-foreset-bottomset structure, suggesting that the tidal range at the time of their deposition was not as great as it is now. Casts and molds of *Portlandia glacialis* (Gray) have been found in this lithosome but, of course, no date is available; 3) A glaciofluvial lithosome consisting of sandy gravel which overlies the two marine lithosomes.

The two marine lithosomes were designated the "Advocate Harbour Member" and the fluvial lithosome the "Saint's Rest Member". The top of the marine member rises from mean sea-level at Saint's Rest to over 135 feet at Advocate Harbour.

Some observations of features indicating emergence of the north shore of Minas Basin were made by this writer prior to the publication of their results by Swift and Borns (1967a, 1967b). The results are not always in agreement, particularly with respect to the amount of emergence indicated. The writer took the top of the terrace as a minimum for marine emergence, when in fact the thickness of the glaciofluvial lithosome ("Saint's Rest

Member") must be subtracted where it is present. At Advocate Harbour the writer obtained a figure of 110 feet as a minimum for emergence. This does not agree well with the 139 feet obtained by Swift and Borns (1967b, fig. 14), but it should be stressed that the writer's measurement is expressed as a minimum.

The evidence for emergence in the Advocate Harbour area consists of a series of ridges cut into or developed on outwash deposits. Swift and Borns (1967b, p. 697) recognized "a series of ridges which resemble the modern spits. These enclose an emerged lagoon backed by a wave-cut bluff similar to that behind the modern Advocate lagoon". It was for the base of this bluff that the writer obtained a height of 110 feet (using an altimeter with a bench mark as datum), and from this the minimum for marine emergence was obtained.

Between Spencers Island and Parrsborough Shore the glaciodeltaic lithosome was observed at a number of places, for example in the valleys of Mahoney Brook and Fowler Brook. But between Parrsborough Shore and Wards Brook the terrace is not represented at all because here the coast coincides with the southern edge of the Cobequid Mountains. It is possible that none of the deposits which make up the terrace to east and west were deposited in this area, but it seems more likely that they were deposited and have since been removed because of the lack of a protecting sedimentary upland to the south.

The terrace appears again at the west end of the settlement of Wards Brook where a small stream cuts through the terrace producing cliffs in which bands of red clay are exposed at heights up to 47 feet above present high-water level. The clays are thought to be the foreset beds of the glaciodeltaic lithosome of Swift and Borns (1967b). If this is the case, then water was a minimum of 47 feet higher than present high-water level at some time since deglaciation. Unfortunately, no shells were found in the clay and absolute dating is not possible.

In the Wards Brook - Fox River area the height of the surface of the terrace was measured at three different places and found to be 97 feet, 97 feet and 90 feet. These heights correspond closely with the height (30 meters) for the "Saint's Rest Member" (glaciofluvial) obtained by Swift and Borns (1967b, Fig. 15) for this area. This is not the figure for marine emergence because the thickness of the "Saint's Rest Member" has to be subtracted; by doing this, Swift and Borns obtained a figure of 24 meters (79 feet).

Between Wards Brook and Parrsboro three prominent characteristics of the terrace can be seen easily in the field and also on air photographs. First, the lower part of the Fox River, where it cuts across the "Five Islands Formation", has produced a very well-marked flight of terraces, formed when sea-level was higher than at present. Some of the terraces are truncated by the present cliffs to the west of Fox Point (Fig. 89) rather than merging into them, suggesting that there has been considerable retreat of the cliffs in this area since the formation of the terraces. Second, the surface is pitted by numerous kettles, some of which are occupied by lakes, while others have been filled in by organic and inorganic materials. In two places, one east of Port Greville and the other on the east side of the mouth of Ramshead River, the hollows have been truncated by coastal erosion. The first one mentioned is close to the truncated river terraces of the Fox River. Here about half the kettle seems to have been removed. Third, the south face of the Cobequid Mountains forms a sharply -defined northern edge to the terrace and has some of the superficial characteristics of an uplifted cliff line. However, the base of the scarp is usually over 150 feet above sea-level and it is extremely doubtful whether during late-glacial or post-glacial times sea-level ever reached this height and, consequently, as far inland as the Cobequid Mountains. This opinion is confirmed by the fact that there is "a relict pattern of braided channels radiating from Parrsboro gap ... indicating that sea level has not been higher than this surface since it formed" (Swift and Borns, 1967b, p. 698). The surface referred to is less than 150 feet above sea-level. As with the Fox River, in the Lower Parrsboro Valley there is a well-developed series of river terraces which deserve detailed study.

East of Parrsboro the terrace is discontinuous and confined to valleys, principally the valley of the Moose River, the valleys of those rivers converging on Lower Five Islands and Five Islands, and the lower part of the Economy River Valley. The altitude of the terrace, which is particularly well-developed in the area of Five Islands and Lower Five Islands, was measured (using an altimeter with a bench mark as datum) at three localities and found to be 48, 63 and 81 feet, indicating that it is by no means flat. On air photographs many small ridges, valleys and kettles can be distinguished, the general arrangement of which - in particular the fact that the ridges bear no relationship to the present coastal trend - indicates that the terrace is not a marine-formed surface. The writer was not able to locate the top of marine-deposited material in this area, but Swift and Borns (1967b, Fig. 15) gave a figure of 15 meters (50 feet) so at least one of the heights measured by the writer may have been on the surface of the glaciomarine lithosome.

Between Five Islands and Saints' Rest patches of the "Five Islands Formation" form the surface, but in all cases it is the glaciofluvial lithosome that is exposed. As mentioned earlier, it is characterized by an irregular, sometimes hummocky surface, in some cases pitted with kettles. Where they reach the coast, some of the kettles are gradually being eroded. According to Swift and Borns (1967b) the glaciomarine lithosomes descend to mean sea-level at Saints' Rest.

Although the surface of the terrace between Five Islands and Saints' Rest does not represent the amount of emergence, there is morphological evidence that emergence has occurred. This takes the form of river terraces and abandoned meander scrolls in the lower parts of the rivers flowing into Minas Basin; namely, Five Islands River, East River of Five Islands, Economy River, and Bass River.

Between Saints' Rest and Truro no clear evidence of emergence exists. There is a low-lying plain between the south side of the Cobequid Mountains and Cobequid Bay, but as far as the writer can tell there is no evidence that the materials of which it is composed were deposited under marine conditions, or were submerged and reworked by wave action. The irregular topography suggests that the plain is underlain by outwash material. In places the soft Triassic bedrock appears at the surface and has been moulded into drumlinoid forms.

Truro stands about 60 feet above sea-level on a gravel plain which was thought by some (Goldthwait, 1924; Stevenson, 1958) to be of deltaic origin. There are few good exposures and the writer was unable to determine whether the plain is of deltaic origin or not. If it is, this indicates emergence of 60 feet for the Truro area, which is at variance with the conclusion of Swift and Borns (1967b) that the zero isobase passes through Saints' Rest, well to the west of Truro (Fig. 69).

Although there is no direct proof that the Truro plain is not of deltaic origin, there is some circumstantial evidence; for example, there is nothing to suggest an emergence of the south shore of Cobequid Bay between Truro and the mouth of the Shubenacadie River. Along this stretch of coast there are numerous drumlin-shaped landforms some of which have been partially eroded by the waters of Cobequid Bay. It can be seen that they are composed of soft Triassic sandstone: they are "rock drumlins" or "false drumlins". The sandstone of which they are composed is very easily eroded when it is exposed along the present coast; therefore, if after the formation of the rock drumlins sea-level had been any higher than it is at present, it is to be expected that some record in the form of cliffs, terraces or notches would have been preserved. No such indications were found.

Another piece of circumstantial evidence against the Truro plain being of deltaic origin is that no marine deposits are found in the Shubenacadie Valley (Hughes, 1957). It is possible, of course, that ice continued to occupy the Shubenacadie Valley after it had withdrawn from Cobequid Bay, allowing marine deposition in the east end of Cobequid Bay but not in Shubenacadie Valley. However, Hughes (1957) thought almost the reverse occurred: that Cobequid Bay and the lower reaches of Shubenacadie River became ice free before the interior of Nova Scotia.

One thing that requires explanation is the steepness of the lower part of the Shubenacadie Valley (Fig. 29). The general, overall form of the valley is thought to pre-date at least the last glaciation (and possibly the whole Pleistocene) although it almost certainly formed a line of ice movement and as a result was modified in detail.

There is very little evidence of emergence along the estuary of the Avon River. One half-mile south of Upper Burlington, on the north bank of the Kennetcook River, a gravel quarry is cut into gravel which may have been deposited when the Kennetcook flowed at a higher level than now; but the top of the gravel is only about 12 feet above present high-water level, so the amount of emergence indicated is small. On the west side of the Avon estuary, at Hantsport, is another exposure of gravel which may have been deposited when the river flowed at a higher level. The emergence indicated is more than 25 feet but less that 50 feet. And, finally in this area, at Avonport Station clay which is very similar in appearence to that being deposited in the inter-tidal zone today is exposed in two clay pits. However, on inspection of the topography of the area on air photographs, the pits were found to be cut into rounded drumlin-shaped forms and the clay may be of glacial rather than marine origin; no shells were found. The drumlinoid forms show no evidence of benching or levelling by the sea.

There is some evidence for emergence of the north end of the Annapolis -Cornwallis Valley but it is not always very clear. Some things may be taken to be of marine origin, but usually there are equally plausible alternative explanations for them. In the Gaspereau Valley, one-fifth of a mile north of Wallbrook, glaciofluvial sands and gravels which dip north-west towards the Gaspereau are exposed. On top is a layer of almost horizontal sand and gravel which might have been formed when the water level stood higher in the Gaspereau Valley. One piece of evidence that supports this is that, three-tenths of a mile to the south-east, and at a greater height, there is another gravel exposure with the steeply dipping glaciofluvial deposits, but without the horizontal "marine" deposits. It is suggested that, when water level was higher than at present in the Gaspereau Valley, it overtopped the glaciofluvial deposits at the lower site but not at the higher one. There is no morphological evidence as to where the shoreline might have been and the most precise figure that can be given for emergence is over 50 feet (the height of the gravel layers at the first site) and below 100 feet (the altitude of the surface of the second exposure).

In the Cornwallis Valley a somewhat similar situation exists about a mile south-west of Port Williams. Here, on the north side of the Cornwallis River, at least 50 feet of surficial material which lies in irregular but definite bands and consists principally of red sand with a few narrow bands of red clay is exposed. In appearence it is similar to the sands and silts being deposited in the inter-tidal zone of Minas Basin at the present; on this basis it is suggested that it was deposited when the water in Cornwallis River was up to a minimum of 50 feet higher than at present. No shells were found in the clay but this is not surprising because at the present few shell fish manage to survive in the sediment -laden waters of Cobequid Bay and Minas Basin.

At Wolfville, Coldwell (1896, p. 173) found "good evidence ... that the sea was at one time nearly 50 feet higher than at present in an old beach formation extending along the line of Acadia Street, parallel to the present water frontage." The writer could find nothing to contradict or to verify this opinion, although the height given by Coldwell coincides fairly well with the height given earlier for a point a few miles up the Cornwallis River. MacNeil(1951a, p. 48) listed a "raised beach" at 80 feet at Wolfville. Further inland, north of Coldbrook Station, Churchill (1918-

1919) recognized what he described as an "abandoned marine sand bar" with its surface at 50 feet above sea-level.

On the north side of the Cornwallis Lowland there is a west-south -west/east-north-east trending valley, in part occupied by Pereau Creek. At Woodside, on the north side of the valley, two ridges run parallel to the trend of the valley; these were thought by Churchill (1918-1919) and Crosby (1962) to be of marine origin. The more prominent of the two ridges has an altitude just in excess of 75 feet[1] and is composed principally of red sand. Its alignment and composition suggest a marine origin: if this is the case, sea-level was once at least 75 feet higher in post-glacial time than it is at present. MacNeil (1951a, p. 48) listed a "raised beach" at Woodside at 100 feet.

Between the mouth of Pereau Creek and Borden Brook the coast is cut into the non-resistant Blomidon Shale of the Annapolis Formation and behind the coast rises the North Mountain escarpment, capped by basalt, which eventually forms the coast north of Borden Brook. As the shale is easily eroded, the cliffs are retreating rapidly and in places are nearly vertical, being up to 100 feet high. Between the cliffs and the face of North Mountain is an undulating plain, much of which lies between 75 and 100 feet above sea-level. The mounds and hollows on the plain result either from the moulding of the bedrock by ice or from irregular glacial deposition. There is no alignment of ridges or mounds that would suggest that the sea has washed over this plain and redistributed sediment. Moreover, as the shale is so easily eroded, it is to be expected that even a short period of

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[1] The height was taken from the 1:50,000 map, 21H/1W. The datum for this and all other topographic maps referred to in this thesis is Canadian Geodetic Datum.

still-stand of the sea on the plain would leave some erosional evidence: there is none. The conclusion reached is that sea-level has not stood as high as 75 feet above the present level since the plain was moulded into its present form. However, MacNeil (1951a, p. 43) stated that there is an "old beach" at about 200 feet at Lower Blomidon.

There is no very clear evidence of emergence along the south coast of Minas Channel as far south as Black Rock Lighthouse, although on the north side of Scots Bay is what might be an old cliff. Set 150 to 225 feet back from the contemporary cliff, it is about 20 feet high and has an accumulation of boulders at its base, which is about 50 feet above sea-level. MacNeil (1951a, p. 48) listed "raised beaches" up to a height of 200 feet at Scots Bay but gave no reason for thinking them to be of marine origin.

Despite the fact that Purdy (1951, p. 41) listed "raised beaches" at: Baxter's Harbour (100 feet), Long Beach (40 feet), Shoal Point (30 feet), Huntington Point (90 feet), and Chipman Brook (60 feet) and that Hudgins (1960, p. 176) listed another at Hall's Harbour (70 feet), the writer found no conclusive evidence of emergence between Scots Bay and Canada Creek. A possible reason for the lack of evidence is the steepness and height - often over 100 feet - of cliffs along this stretch of coast. These cliffs are cut into resistant basalt and almost certainly pre-date the last glaciation. Thus, if following the last glaciation relative sea -level rose to 50 or 75 feet above its present level, water would have rested against an almost vertical cliff where it could not form a wave-cut platform and a new cliff (Fleming, 1965).

At Black Rock Lighthouse, approximately one mile west of Canada Creek, there is an isolated remnant of a wave-cut platform. The remnant, small and extending no more than 300 feet along the coast, appears to have been preserved because the surficial materials into which it is cut are protected by bedrock on the seaward side.

At Harbourville, east of the mouth of Givan Brook, there is a more extensive remnant of an emerged marine platform. After a gap of about a mile and a half, a platform and cliff reappear at Turner Brook. They do not exist between Ogilvie and Meekin Brook perhaps because the cliffs in this area are too high and too steep for any former sea to have left any remnant on or above them.

Further west an abandoned cliff and marine platform are preserved west of Meekin Brook, and another terrace and cliff - not too distinct - east of the settlement of Morden. Between Morden and the mouth of Moody Brook a bench, probably of marine origin, exists, but in this case there is no prominent scarp at the back of it.

The settlement of Margaretsville is built on a promontory, one of the few along the south-east Fundy coast. On both sides of the settlement there is an emerged terrace which appears to be of marine origin. It is cut into or developed on surficial materials and the back of the terrace is marked by a change in the nature of the land surface rather than by a marked cliff line. Whereas the terrace has a relatively even surface, the land behind it is more irregular and rolling in appearance.

At Port George there is a clearly-marked cliff (Fig. 48). Seaward from it the land slopes gently down to the present shoreline. The emerged cliff/platform contact is not easily located because slumping, soil creep, and general downhill movement of loose material since its formation have produced a curved surface between the two slope elements. However, by taking the mid-point of curvature as the contact, it was found that the base of the old cliff is at a height of 51 feet (measured by altimeter, using the debris line as datum).

A mile and a half south-west of Port George at Cottage Cove there is a deposit of sand and gravel, probably of marine origin, the surface of which is at a height greater than 50 feet but less than 100 feet. Finally, at Hampton several low ridges composed of surficial deposits are aligned roughly parallel to the present shore. They may be of marine origin[1].

Hickox (1958) measured the height, above spring high-tide, of the cliff/platform contact at four of the localities mentioned in this account. His results were: Harbourville, 62 feet; Turner Brook, 65 feet; Margaretsville, 66 feet; and Port George, 65 feet (compared with 61 feet obtained by the writer).

The fact that the height of the cliff/platform contact varies little from place to place led Hickox to reason that a single still-stand of the sea is indicated by the four sites he surveyed. If this is so, an isobase drawn through the localities should run parallel to the south Fundy shore, which agrees closely with most "isobase" maps that have been drawn for the area (Figs. 61-69). Hickox believed that the tidal range at the time of formation of the "strandline" was the same as it is today and that emergence of about 65 feet is indicated. He concluded that this was the highest post-glacial sea-level. For reasons explained earlier (p.108) the writer doubts that the tidal amplitude was as great as it is today. An uplifted cliff/platform contact is now 65 feet above the contemporary

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[1]Hudgins (1960, p. 176) reported a "raised beach" at an altitude of 60 feet at Hampton.

cliff/platform contact, but even if both were formed at high-tide level it would be incorrect to conclude that mean sea-level has fallen by 65 feet (Fig. 49).

Nothing is found between Hampton and Digby Gut, although in some respects this stretch of coast would seem to be ideal for the preservation of remnants of former marine features. The land slopes gently from the crest of North Mountain towards the sea to the north-west and it might be expected that if sea-level was higher in the past it would have cut notches into the gentle slopes. However, the slope of the land may have prevented the formation of any erosional landforms. Fleming (1965, p. 803) stated that "the maximum rate of mechanical erosion by waves will occur on initial land slopes between 1/4 and 1/10." The land slopes here are usually less than 1/10. As the land is underlain by resistant basalt with only occasional glacial and glaciofluvial covering, it is probable that, even if sea-level were once higher, regression took place without leaving any erosional impact on the landform. The absence of marine depositional forms is harder to explain, except that perhaps there was a lack of raw materials from which constructional forms could be built. It is notable that all the indicators of higher sea-levels along the Fundy side of North Mountain have been cut into or preserved on surficial deposits.

Changes in base level may be reflected in the long profile of a river. On the Fundy side of North Mountain many small streams, normally less than 4 miles in length, flow in a north-westerly direction. Long profiles of each of these streams (134 in all), some of which are shown in Figure 90, were constructed to see whether there is any indication of changes in base level. The most noticeable feature is that several of the streams between Wheaton Vault and Sheep Shearer Brook show a steepening at 100 feet or 50 feet and in some cases the drop to the shore is vertical. It is possible that this is a result of a drop in base level of between 50 and 100 feet and that, since this occurred, the streams have not had time to adjust to the new base level. However, the fact that the streams north of Wheaton Vault and those south of Sheep Shearer Brook do not show the steepening appears to contradict this hypothesis.

It seems possible that during post-glacial time the Annapolis-Cornwallis Valley was an arm of the sea and that North Mountain was an island. However, Caribou Bog, situated at the height of land between the headwaters of the Annapolis River and those of the Cornwallis River, is 125 feet above sea -level; therefore, submergence of more than 125 feet would have been needed, assuming there has been no tilting along the axis of the lowland. An isobase drawn through the four sites levelled by Hickox runs parallel to the Annapolis-Cornwallis Lowland, which suggests that there has been little or no tilting along the axis of the lowland.

It has been shown that there is evidence for emergence of between 60 and 70 feet along the Fundy coast and that in general the amount of emergence decreases to the south-east. Therefore, on the basis of information so far cited, there is no reason to believe that the lowland was completely flooded in post-glacial time.

However, there is some basis for believing that sea-level was once higher in Annapolis Basin than at present; for example, Bailey (1898, pp. 20M-21M) stated that "the brick clays of Middleton [which was near the tidal limit before a dam was constructed at Annapolis Royal] contain layers filled with marine shells, together with remains of star-fishes *Ophiopholis*". Hickox (1962a, p. 30) reported that the brickyards from which the fossils were taken is 47 feet above mean sea-level and that the top of the clay at Middleton railway station is 57 feet above mean-tide. Therefore, within the Annapolis Valley clear evidence exists of a minimum of 47 feet, and probably 57 feet, of emergence.

Goldthwait (1924, p. 151) reported a "beach-like slope" at 85 feet (datum not given) north of Digby; a "rocky cliff about 80 feet above high tide" at Prim Point;and three low ridges resembling beaches at 88, 73 and 43 feet (data not given) east of Prim Point. In no case is the evidence for the marine origin for these features very convincing, so the writer is prepared to specify only that there is a possible 88 feet of emergence at the south end of Annapolis Basin.

Swayne (1952) mapped "abandoned shorelines" on the north side of Annapolis Basin from Victoria Beach to a point two miles north-east of Kinsdale; on the south side of the Basin from Clementsport to Smiths Cove Station; and along the west side from Digby to Prim Point, their heights ranging up to 165 feet at Clementsport. No justification is given for the use of the term "abandoned shoreline" and, as the "abandoned shoreline" north of Annapolis Basin coincides almost exactly with the lithological boundary between the non-resistant Annapolis Formation below and the resistant North Mountain Basalt above, the marine origin of the features is in doubt.

Annapolis Basin was not studied in detail by the writer, but three sites were investigated to see whether they displayed evidence of emergence. North-east of Round Hill there is an extensive exposure of surficial material composed of well-layered sand and fine gravel. The layers dip in a northerly direction and at amounts up to 25°. On the basis of these facts the deposit is thought to be of glaciofluvial origin, laid down by water issuing from ice on South Mountain. There is certainly nothing to suggest that it is of marine origin.

At Imbertville, on the north side of the mouth of Roach Brook, there is an exposure of layered gravel. The layers dip towards the south, towards the nearest water, and may have been formed along a shore when sea-level was higher than at present. The altitude of the top of the gravel layers is 90 feet (measured by altimeter from the nearest bench mark) which agrees very closely with 88 feet above mean-tide quoted by Goldthwait (1924, p. 152) for the highest gravel terrace in the valley of the Bear River, one and a half miles east of Imbertville. Goldthwait regarded the terrace as a remnant of an old floodplain and therefore as an indication of the amount of emergence around Annapolis Basin.

South of The Joggins, the surficial deposits exposed consist of layered sand and gravel and are, therefore, water deposited. In this case the dip is gentle - 6^{o} was the maximum measured - to the south-south-east towards Acacia Brook which occupies a steep-sided valley and flows into the south end of Annapolis Basin. The top of the gravel is at an altitude of 106 feet (datum a bench mark). The sand and gravel may have been deposited when sea-level was higher in Annapolis Valley, but there is no supporting evidence: no cliff, no wave-cut platform, no clay, and no fossils. Therefore, the marine origin of the deposit is not certain.

On the basis of the foregoing discussion about the Annapolis Basin and Valley the most definite statement that can be made is that there was submergence of at least 57 feet and perhaps as much as 90 feet during post -glacial time, insufficient for the complete flooding of the Annapolis -Cornwallis Lowland.

The question of the emergence of the mainland coast of St. Marys Bay has received some attention from previous workers. Goldthwait (1924, p. 152) reasoned that a gravel terrace at 85 feet above mean-tide at the mouth of the Sissiboo River marks the old sea-level at that point. Swayne (1952, p. 56) gave the height of a number of "raised beaches" at Brighton and Barton ranging up to 200 feet, but the marine origin of these features is doubtful. Dunlop (1952) mapped two emerged deltas south of Belliveau Cove and at Little Brook, both of which appear to be at altitudes of less than 50 feet.

Marine shells (*Mercenaria mercenaria*) were found at Gilbert Cove at an altitude of 46 feet (measured by altimeter from a bench mark). They were well-preserved but irregularly distributed in a gray stony clay which was covered by sand and gravel before excavation took place at the site. The shells were dated by the Geological Survey of Canada (GSC - 887) and found to be greater than 39,000 years of age. Medcof, Clarke and Erskine (1965, p. 631) claimed the *Mercenaria mercenaria* is not found today in the cool-water areas of Canada's east coast. It is possible, therefore, that the shells at Gilbert Cove were deposited during a high stand of the sea before the last glaciation when the water may have been warmer than at present. However, shells of similar age (GSC - 695; >39,000) were found by Grant one mile south-east of Cape St. Mary[1]. Here they were found at an altitude of 15 feet -35 feet in a stony clay which is overlain by glaciofluvial gravel. Grant interpreted the shell-bearing stony clay as till, the shells being picked up from the sea bottom. The similarity between the situation at Gilbert Cove and that at Cape St. Mary suggests

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[1]For details see Lowdon and Blake Jr. (1968), *Geological Survey of Canada Radiocarbon dates VII*, Geological Survey of Canada Paper 68-2, Part B, page 211, 1968.

that the shells found at the two places may have been emplaced in the same way. If Grant's explanation is correct, then the height at which the shells are found gives no indication of the amount of emergence. However, Grant, in his description of the stratigraphic arrangement at Cape St. Mary, stated that the glaciofluvial gravel above the shell-bearing clay has been worked by post-glacial marine action up to 40 to 45 feet. There are indications elsewhere that emergence of at least this amount has taken place.

The settlements of Comeauville and Saulnierville are built on a ridge, just over 50 feet high, which runs parallel to the shore for a distance of 3 miles and is broken only where the Leblanc Brook cuts through to the sea. It is thought that this ridge is of marine origin and indicates emergence of at least 50 feet. Although the writer has no field evidence to support this view, there are several factors that support the conclusion: 1) The underlying bedrock is the Goldenville Formation (Early Ordovician or Late Cambrian) which consists of graywacke, minor argillite, shale, and mica schist (Taylor, 1969). The bedding planes north of the mouth of Leblanc Brook are nearly parallel to the coast, but to the south they are at an angle to it. In view of this, it seems unlikely that the ridge, which for its whole length is parallel to the coast, is a bedrock feature. 2) Inland from the ridge there is a lowland occupied by Leblanc Brook and Freshwater Stream. It is marshy and would have been a lagoon if sea-level had been 50 feet higher than now. Behind the "lagoon" are steep slopes which probably represent the old coast. 3) When viewed on air photographs, the ridge can be seen to have a series of minor lineations parallel to it, particularly on the seaward side. 4) On a soil survey map of Digby County the soil on the ridge is shown to belong to the Digby Series which has "coarse gravelly parent material" underlain by "a compact, relatively impervious bed of marine sediments" (Hilchey, Cann and MacDougall, 1962, p. 37).

Although the ridge-like landform is not preserved, the Digby Soil Series continues south to Meteghan. At Meteghan Centre, at a height just below 50 feet, an exposure reveals layers of sand and gravel dipping towards St. Marys Bay. In appearence they are typical of marine deposits.

Further south at Mavillette there are more patches of the Digby Soil Series. The sand and gravel on which they are developed is exposed on either side of a small intermittent stream which flows south-west towards the Atlantic Ocean through a marsh-covered lowland which was once an embayment of the coast. The embayment is now cut off from the open sea by sand spits. On its east side layers of sand and fine gravel are exposed at an estimated height of 30 feet above high-water level; this can be taken as a minimum for emergence.

The area between the northern end of St. Marys Bay and the southern end of Annapolis Basin is similar to the Annapolis-Cornwallis Lowland in that it seems possible that it was completely submerged in post-glacial time. However, the height of land between the two water bodies is over 100 feet above sea-level and there is no evidence that sea-level has been that high in the area in post-glacial time. Swayne (1952) mapped an "abandoned shoreline" between Rossway and Roxville at a height of over 100 feet but, like the "abandoned shoreline" along the north side of Annapolis Basin, it coincides almost exactly with a lithological boundary and is thought to be of subaerial rather than marine origin.

The figures obtained for marine emergence for Annapolis Basin and for the mainland coast of St. Marys Bay are less than 100 feet, with the possible exception of the deposit south of The Joggins. Therefore, it is concluded that the two water bodies have not been joined in post-glacial time.

Some evidence exists for the emergence of Digby Neck, Long Island, and Brier Island. At a point on the east side of Gulliver Cove, at the northern end of Digby Neck, cliffs approximately 25 feet high are composed of boulder clay below, with a coarse boulder bed, about three feet thick, at the top. The boulder clay consists of a great size range of pebbles and boulders set in a red-clay matrix. The boulder bed at the top probably represents the result of washing out of fine material when sea-level was higher than at present.

South of Gulliver Cove a deep, fault-determined valley cuts through the basalt of Digby Neck. It is drained by the Cassaboom Brook which flows north to Gulliver Cove and by Haight Brook which flows south to St. Marys Bay. The height of the land between the two is greater than 50 feet but less than 100 feet. Sea-level rise of 100 feet would result in St. Marys Bay and the Bay of Fundy being connected by a strait similar to Petit Passage and Grand Passage further south, but there is no evidence that submergence of this amount did occur.

A somewhat similar situation occurs at Centreville, further south on Digby Neck, except that here the gap does not extend completely across the peninsula. It is partly filled by glacial deposits which have been wave-washed, resulting in the removal of fine material and the leaving behind near the surface of coarse material.

At Sandy Cove there is yet another gap through the basalt ridge, the north-western part of the gap being occupied by the waters of Sandy

Cove and the south-eastern part by East Sandy Cove. The area between these two water bodies is occupied by a complex system of surficial deposits. Goldthwait (1924) thought the system was a delta which indicated 125 feet of emergence. He reported the occurrence of shells in the deposits but did not find any himself, and the writer has not located any report of shells in the area in the more recent literature. Probably some at least of the deposit is of marine origin, but the situation is not as clear as suggested by Goldthwait.

At the south end of Long Island a mile east of Freeport at between 50 and 100 feet, there are gravel deposits whose topographic form is such that they are thought to represent an uplifted sand and shingle bar. There is also a low scarp, probably of marine origin, on the west side of the cove which occupies the lowland between the two north-east/south-west trending ridges that constitute Long Island. The base of the scarp is below 50 feet.

On the other side of Grand Passage, behind Westport on Brier Island, Goldthwait (1924, p. 149) reported the occurrence of "beaches" up to a height of 115 feet above mean sea-level which he thought was the limit of post-glacial emergence for the area. On air photographs a series of ridges running parallel to the present coast can be seen west of Westport. They are probably of marine origin, but Goldthwait's figure of 115 feet may be high.

Summary

Information about emergence of the Fundy coast as described in this chapter is summarized in Table 2 and Figure 91. The following facts about the table should be noted:

(1) Only places for which a height is available are included.

(2) The figures given are the minimum estimates for the locations cited: emergence may have been much greater at some of the places.

(3) Where two figures are given for one location, they represent either two different estimates by different workers or two sites which are so close together that they cannot be designated by separate place names.

(4) Where possible the datum from which the heights were determined is given. The designation "map" means that the height of the feature indicating emergence was obtained by plotting its location on the relevant 1:50,000 map and by reading its height from the map. This is, of course, imprecise in that the contour interval on most of the maps is 50 feet, although there are a few with a 25 feet interval. Thus, the height of a feature is usually given in the text in the following manner: "above 50 feet but below 100 feet". Since the table deals with minimums this height would be recorded as 50. The datum is stated in order to avoid confusion that might result from equating something which is say 56 feet above mean sea-level with something that is 56 feet above high-tide level.

(5) The heights quoted are those obtained from all types of indicators of emergence.

(6) Where the heading "certain" is used, shells were found in sediments located at the altitude given or in sediments extending up to the altitude given. Nothing is completely certain in geomorphology, but the finding of shells in the conditions described earlier in this chapter is the surest indicator of marine origin.

(7) Where the heading "doubtful" is used, the figure given is usually out of line with figures for adjacent or nearby areas, or there is some very strong reason for doubting the marine origin of the feature whose height is given.

TABLE 2

MINIMUM HEIGHTS OF POST-GLACIAL EMERGENCE AT LOCALITIES AROUND THE BAY OF FUNDY (HEIGHTS IN FEET)[1]

Site Number and Location	Certain	Probable	Possible	Doubtful
St. Stephen-St. Andrews[2]		150 (map)[3] : GADD, 1970b		
1. Benson Corner	85 : GADD[4]			
2. Sand Point	46 (debris line) : WELSTED	100 : GADD[5]		
St. George map area[6]		200 (map) : GADD, 1970b		
3. Hawkins Hill			50- (map) : WELSTED	
4. Pennfield Plain	130 : GADD[7]	250 (map) : GADD, 1970b		
5. Little Lepreau Basin		50 (map) : WELSTED		
6. Dipper Harbour		49 (debris line) : WELSTED		
7. Musquash River		90 (debris line) : WELSTED		
8. Lorneville		200 (map) : WELSTED		
9. Taylor Peninsula-Sheldon Point	59 (cliff base) : WELSTED			
10. Sheldon Point Peninsula		100 (map) : WELSTED		
11. Negrotown Point		50 (cliff base) : WELSTED		
12. One mile north of Taylor Peninsula	100 (map) : WELSTED			
13. Red Head		40 (cliff base) : WELSTED		
14. Silver Falls	100 (map) : WELSTED			
15. Midwood	40 : MELVIN, 1966			
16. Little River		150 : MELVIN, 1966		
17. Lawlor Lake	100 (map) : MATTHEW, 1879			
18. Mispec River			100 (map) : WELSTED	

TABLE 2 - Continued

Site Number and Location	Certain	Probable	Possible	Doubtful
19. Black River-Emerson Creek		50 (map) : WELSTED		
20. Honeycomb Point		155 (debris line) : WELSTED		
21. St. Martins		100 (map) : WELSTED		
22. Macomber Point		122 (debris line) : WELSTED	150 (map) : WELSTED	
23. Mosher River		34 (debris line) : WELSTED		
24. Wolfe River		115 (debris line) : WELSTED		
25. Fundy National Park		135 (spring high-tide) : BORNS and SWIFT, 1966		
26. Alma River		58 (high-tide level)[8]: WELSTED		
27. Alma			50 - (map) : WELSTED	
28. Dennis Beach		78 (debris line) : WELSTED		
29. Little Rocher		24 (debris line) : WELSTED		
30. Cape Enragé		50 (map) : WELSTED		
31. Albert		64 (bench mark) : WELSTED		
32. Advocate Harbour		139 (mean sea-level) : SWIFT and BORNS, 1967b	110 (bench mark) : WELSTED	
33. Wards Brook		47 (bench mark) : WELSTED		
34. Wards Brook - Fox River		79 (mean sea-level) : SWIFT and BORNS, 1967b		
35. Five Islands-Lower Five Islands		49 (mean sea-level) : SWIFT and BORNS, 1967b		
36. Saints' Rest		0 (mean sea-level) : SWIFT and BORNS, 1967b		
37. Truro				60 : GOLDTHWAIT, 1924
38. Upper Burlington			12 (high-tide level)[9] : WELSTED	
39. Hantsport			25 (map) : WELSTED	

TABLE 2 - Continued

Site Number and Location	Certain	Probable	Possible	Doubtful
40. Gaspereau Valley		50 (map) : WELSTED		
41. Port Williams			50 (map) : WELSTED	
42. Wolfville			50 : COLDWELL, 1896	80 : MacNEIL, 1951
43. Coldbrook Station			50 : CHURCHILL, 1918-1919	
44. Woodside			75 (map) : WELSTED	100 : MacNEIL, 1951
45. Lower Blomidon				200 : MacNEIL, 1951
46. Scots Bay			50 (map) : WELSTED	200 : MacNEIL, 1951
47. Baxter's Harbour				100 : PURDY, 1951
48. Long Beach				40 : PURDY, 1951
49. Shoal Point				30 : PURDY, 1951
50. Huntington Point				90 : PURDY, 1951
51. Chipman Brook				60 : PURDY, 1951
52. Hall's Harbour				70 : HUDGINS, 1960
53. Port George		65 (spring high-tide) : HICKOX, 1958 61 (debris line) : WELSTED		
54. Cottage Cove		50 (map) : WELSTED		
55. Hampton			60 : HUDGINS, 1960	
56. Harbourville		62 (spring high-tide) : HICKOX, 1958		
57. Turner Brook		65 (spring high-tide) : HICKOX, 1958		
58. Margaretsville		66 (spring high-tide) : HICKOX, 1958		
59. Middleton	47 (mean sea-level) : HICKOX, 1962[10]	57 (mean sea-level) : HICKOX, 1962		

TABLE 2 - Continued

Site Number and Location	Certain	Probable	Possible	Doubtful
60. Digby			85 : GOLDTHWAIT, 1924	
61. Prim Point			80 : GOLDTHWAIT, 1924	
62. East of Prim Point			88 : GOLDTHWAIT, 1924	
63. Clementsport				165 : SWAYNE, 1952
64. Imbertville		90 (bench mark) : WELSTED		
65. Bear River		88 (mean sea-level) : GOLDTHWAIT, 1924		
66. The Joggins			106 (bench mark) : WELSTED	
67. Sissiboo River		85 (mean sea-level) : GOLDTHWAIT, 1924		
68. Brighton				200 : SWAYNE, 1952
69. Belliveau Cove			50 - : DUNLOP, 1952	
70. Little Brook			50 - : DUNLOP, 1952	
71. Cape St. Mary		40 : GRANT, 1968[11]		
72. Comeauville-Saulnierville		50 (map) : WELSTED		
73. Meteghan Centre		50 - (map) : WELSTED		
74. Mavillette		30 (high-tide level)[12] : WELSTED		
75. Rossway-Roxville				100 : SWAYNE, 1952
76. Gulliver Cove		25 (cliff base) : WELSTED		
77. Sandy Cove			125 : GOLDTHWAIT, 1924	
78. Freeport		50 (map) : WELSTED		
79. South end of Long Island		50 - (map) : WELSTED		

TABLE 7 - Continued

Site Number and Location	Certain	Probable	Possible	Doubtful
80. Westport		115 : GOLDTHWAIT, 1924		

1. The names in capital letters are those of the persons who either identified the location or determined the height given in the table.

2. 6. Not shown on Figure 91.

3. Datum from which the height was obtained. If no datum is indicated it was not stated by the person who determined the height.

4. Height given in Lowdon and Blake Jr. (1970), page 56.

5. Personal communication.

7. Height given in Lowdon and Blake Jr. (1970), page 55.

8. 9. 12. Subject to error resulting from incorrect location of high-tide level.

10. Site first located by Bailey (1898).

11. Height given in Lowdon and Blake Jr. (1968), page 211.

(8) The distinction between "probable" and "possible" is much more subjective and is made taking into consideration the thoroughness with which the locality was studied in the field, impressions gained from air-photo and map interpretation, and the opinions of previous workers in the area. An attempt was made to assess the detail and thoroughness with which previous studies were carried out and also the validity of the reasoning used when a feature was labelled "marine".

An assessment of the reliability of the evidence for emergence is attempted because "emerged shorelines" are like erosion surfaces in that, with more detailed study of an area, their number tends to proliferate to a stage where one is recorded at almost every foot from 0 up to the highest level recorded. In some cases the evidence is conclusive whereas in others heights given involve nothing more than speculation. Table 2 is an attempt to record this fact.

The foregoing description of the evidence for emergence around the Bay of Fundy and the summary of this evidence in Table 2 and Figure 91 reveal a number of interesting facts about the area. The most obvious is the scarcity of post-glacial shells from the Nova Scotia coast,which is reflected in the existence of only one "certain" indicator of emergence - at Middleton. Shell casts and moulds have been found in the Five Islands Formation (Swift and Borns, 1967b) but this does not help in determining its precise age; although shells have been reported at Sandy Cove (Goldthwait, 1924), none appear to have been collected in the last 50 years; and those found at Gilbert Cove and Cape St. Mary are much too old to be of post-glacial origin. In contrast, shells are relatively abundant on the New Brunswick side of the Bay south of Mispec, and are even more abundant further south along the coast of Maine (Bloom, 1963; Borns, 1963).

On the whole, the figures for emergence are greater on the New Brunswick side of the Bay than along the Nova Scotia coast. The highest for New Brunswick is Pennfield Plain (250 feet) and for Nova Scotia, Advocate Harbour (137 feet). It is of significance that Advocate Harbour is nearer to the New Brunswick side of the Bay of Fundy than any other place in Nova Scotia where there is evidence of emergence. In general, emergence increases in a north-west direction.

There is considerable variation in the figures given and no apparent pattern emerges apart from that just mentioned. It should be remembered, however, that the figures are for scattered indicators of emergence and that in all probability the upper limit of marine deposition and erosion was influenced by the presence of ice. The figures may indicate the marine limit in some cases, but in other cases they clearly represent something less. Therefore, there is no basis for drawing isolines on the marine limit, and, as there are few accurately heighted and dated sites, drawing of isobases is at this time impossible.

Another notable feature is that there are two negative areas as far as evidence of emergence is concerned: 1) Shepody Bay, Cumberland Basin, and the Nova Scotia side of Chignecto Bay; and 2) the north shore of Cobequid Bay from Saints' Rest to Truro, and the south shore of Cobequid Bay and Minas Basin as far west as the mouth of the Avon River.

Finally, with very few exceptions, emerged erosional marine features are absent from areas underlain by the resistant pre-Triassic rocks and the Triassic basalts which form much of the Fundy coast. Where erosional landforms are preserved, they are cut mainly into the Triassic rocks other than the North Mountain Basalt or into surficial deposits. The conclusion is that, since deglaciation, sea-level has not remained the same long enough for any significant erosion of the resistant rocks. Where there are steep cliffs in resistant rocks, as for example on the west coast of Grand Manan Island and along parts of the Fundy shore of North Mountain, they must have originated in pre-glacial time. Glaciation and sea-level changes probably played a part in steepening the cliffs.

Deglaciation and Emergence

The question arises as to when the Bay of Fundy was ice free and how the time of ice dissipation relates to emergence. It has been shown by King (1969) that ice at its maximum reached the Scotian Shelf. If it is assumed that ice retreated in a north-westerly direction across Nova Scotia, then radiocarbon dates suggest that it had reached the coast of southern New Brunswick (Saint John and southward) by about 13,000 BP [I (GSC) - 7; GSC - 882; GSC - 965] and that the Bay of Fundy was open by this date. Although the Bay may have been open earlier, as indicated by the date for Pennfield Plain, 16,5000 ± 370 (GSC - 1063), it would be unwise to rely too heavily on one date. An end moraine complex, extending along the coast of Maine and projected into New Brunswick (Borns, 1968; Gadd, 1969), may represent an episode in the retreat of the ice sheet. If one thinks in the simplest terms of a steady north-west retreat of the entire ice front, the whole length of the New Brunswick Fundy coast (with the exception of Passamaquoddy Bay) would have been ice free at the latest by about 13,000 BP (Fig. 92). It has been shown that at Sand Point sea-level was a minimum of 46 feet above present high-tide level 12,300 years ago (GSC - 886); and at Pennfield, at 130 feet, 13,000 years ago (GSC-882). If the "isobase" maps (Figs. 61-69) are at all reliable, there has been no appreciable tilting along the axis of the Bay. Therefore, at the head of the Bay, relative sea -level 13,000 years ago would have been at least 130 feet higher than at present and 12,300 years ago, at least 85 feet higher than at present. Consequently Chignecto Isthmus would have been flooded and the tidal range would have been less than it is today.

The sequence of events just described is, in the writer's opinion, unlikely; but it is clear that the coast of <u>southern</u> New Brunswick, with the exception of Passamaquoddy Bay, was free of ice by about 13,000 BP. At that time sea-level was rising at an average rate of 25 feet per thousand years (Shepard, 1963b). It seems likely that as sea-level rose it advanced into the Fundy trough, splitting "New Brunswick ice" from "Nova Scotia ice", at least in the south, so that at the time when shells were deposited at Pennfield and Saint John the ice front had the outline shown in Figure 93. This is similar to the situation suggested by Prest and Grant (1969, Fig. 4) and to that shown on *Retreat of Wisconsin and Recent Ice in North America* (Geological Survey of Canada, Map 1257A)[1]. The tidal fluctuation under these conditions would have been less than at present, though perhaps more than if the Bay were a straight.

Therefore, any uplifted marine feature that can be dated at 13,000 BP or earlier must have been formed under different tidal conditions from those that exist today. How long it would have taken for the ice front to be "pushed" back to, and possibly across, the Chignecto Isthmus is not known although Borns (1966) claimed that about 10,600 years ago, at the time of the occupation of a "Paleo-Indian" site at Debert, all the continental ice

[1]This sequence does not take into account the anomalous date of 16,500 ± 370 obtained for Pennfield Ridge.

had dissipated from northern Nova Scotia. A substantial time-gap exists between 13,000 and 10,600 but Geological Survey of Canada Map 1257A gives the impression that Chignecto Isthmus was free of ice well before 12,500 BP, and an isochrone map drawn by Bryson *et al.* (1969) shows the Isthmus free of ice between 11,000 and 12,000 years BP. Taking the Bay of Fundy as a whole the time available for "Up" in Andrews' (1968a, 1970) formula is probably between 13,000 and 12,000 radiocarbon years.

Evidence of Submergence

The term submergence is used to indicate the downward movement of land relative to sea-level. There is abundant evidence that the level of the sea was once lower relative to the land than it is now and that it has risen, resulting in the drowning of features which were once above sea-level. A distinction must be made between the large scale changes of the Pleistocene and the much smaller variations that have taken place during the Recent Epoch; that is, post-glacial changes.

Some idea of the amount of sea-level lowering during the Pleistocene, and therefore the amount of submergence that has taken place in order to reach the present situation, can be obtained by studying the curves for eustatic changes (Figs. 50 - 60) and the table of sea-level lowering during the last glaciation (Table 1). Even taking the smallest estimate shown in the table, 295 feet by Kuenen (1954), almost all of the Bay of Fundy north of a line joining Point Lepreau and Digby Gut and much of the Gulf of Maine would be above sea-level (*Canadian Hydrographic Chart 4011*). Taking McFarlan's (1961) estimate, 450 feet, all the Bay of Fundy except for a few basins at its mouth would be dry and the Gulf of Maine would consist of a series of water-filled basins. Georges Bank and Browns Bank would be above sea-level but separated by Northeast Channel (assuming the topography of the time of lower sea-level to be the same as that of today).

These are, however, estimates for world-wide changes and do not take local conditions into consideration. Because of the complications resulting from ice cover, isostatic effects (of ice and water), and deposition and redistribution of sediments, there is no guarantee that, even if it could be proved conclusively that eustatic level was -295 feet during the last Wisconsin ice advance, a point in the Bay of Fundy now 290 feet below sea-level would have stood 5 feet above sea-level.

It is necessary, therefore, to search for local evidence of lower sea-levels and in this context it is intended to include not only the Bay of Fundy and its coast, but also the Gulf of Maine, the banks on the continental shelf, and Northumberland Strait.

Coastal Configuration

An indication that submergence has taken place can sometimes be obtained from the coastal outline. South of the study area, Cobscook Bay in Northern Maine has the intricate outline of the classic "shoreline of submergence" (Johnson, 1925) and Passamaquoddy Bay and Letang Harbour in south-west New Brunswick are almost as good examples. The parallel bays and lakes of the lower Saint John Valley (Kennebecasis Bay, Long Reach and Belleisle Bay, Washademoak Lake, and Grand Lake) have the appearence of drowned valleys.

It may be of significance that the area between Passamaquoddy Bay and Saint John Harbour has the highest figure for emergence recorded in Table 2, perhaps indicating that emergence is more nearly complete there than in the areas on either side. This may result from the fact that ice occupied Passamaquoddy Bay and the Lower Saint John Valley after it had disappeared from the area in between. There is some evidence to support this contention in that the radiocarbon date, GSC - 1063,may indicate that ice had withdrawn from the Pennfield Plain by 16,500 ± 370 BP, whereas other radiocarbon dates indicate that Passamaquoddy Bay was not free of ice until about 12,300 years ago and Saint John Harbour, not until about 13,200 BP. Andrews (1968a) argued the case that for sites with similar values for ice thickness - which is equated with distance from the ice edge to each site - but with different dates of delgaciation, the site which was ice-free last will have the smallest post-glacial uplift. In the case of the three sites (Saint John, Pennfield, and Passamaquoddy Bay) in south-western New Brunswick, the distance to the ice edge, which was probably on the Scotian Shelf, was approximately the same. This being the case, it is to be expected that post-glacial uplift will be less in Passamaquoddy Bay and in the Lower Saint John Valley that at Pennfield; this may account for their having a "more" submerged appearence" than the area between them. The writer does not wish to make too much of this point because post-glacial uplift has not been accurately determined and the Pennfield date is anomalous when compared with all other dates for the New Brunswick coast (Appendix 2). Nevertheless, it is a line of study worth pursuing and further radiocarbon dates should help to clarify the situation.

Between Saint John Harbour and Chignecto Bay the New Brunswick coast does not have the outline usually associated with submergence. However, it should not be taken that an intricate coastal outline automatically indicates submergence whereas a straight one does not. If a coast is straight and has high cliffs, a rise in sea-level will not substantially alter its outline, unless the rise is great enough to overtop the cliff. Therefore, the relative straightness of the New Brunswick coast north of Saint John does not mean that submergence has not taken place. The same argument applies to the coast on the opposite side of the Bay - that developed along the north-west side of North Mountain.

At the head of the Bay of Fundy the long estuaries leading into Shepody Bay, Cumberland Basin, and Minas Basin are features usually associated with submergence, and at the southern end of the Bay the intricate outline south of Cape St. Mary is similar to that in the Cobscook Bay - Passamaquoddy Bay area.

The whole southern and south-eastern coast of Nova Scotia consists of a series of bays, inlets, peninsulas and islands - a situation usually associated with submergence. Maps showing isolines of recent crustal movement (based on tide-gauge data) support this interpretation in that the line of zero movement is placed north-west of Nova Scotia's Atlantic coast with negative values to the south-east and positive values to the north -west. The exact position of the zero line varies from worker to worker: Gutenberg (1941) placed it almost exactly along the Fundy coast of Nova Scotia; Moore (1948) showed it passing through Saint John, almost parallel to the New Brunswick coast; and Valentin (1954) showed it slightly north -west of the New Brunswick coast. Whichever is taken as being most nearly correct, the Atlantic coast has a negative value whereas the Fundy coast has a smaller negative value or a positive value. The difference may be due to differential tilting resulting from continuing isostatic readjustment - Urr in Andrews' (1968a) formula . There is also the complicating factor of eustatic changes which will be considered in more detail later.

River-Bottom Profiles

The depth of the bedrock surface at the mouth of a river channel may indicate the base level (sea-level) that prevailed in the past. The depth to the bedrock valley may be taken as a rough indication of the rise in sea-level that has taken place. There are, however, complicating factors such as the possibility that the bedrock floor has been deepened by ice erosion. It is also difficult to fix with any precision the time at which the lower base level existed.

The submarine topography and geology of Passamaquoddy Bay have been subjected to detailed investigation because of the possibility of locating a tidal power scheme in the area (Smith, 1958). In Western Passage, between Deer Island and the coast of Maine (Fig. 70), the bedrock surface is about 400 feet below mean sea-level; and in Head Harbour Passage, between Deer Island and Campobello Island, it is below -400 feet (Upson and Spencer, 1964). Western Passage, the natural outlet for the St. Croix River, is aligned with the suggested general movement of ice along the Oak Bay - St. Croix Valley. Head Harbour Passage, however, is at an angle to this line and Upson and Spencer (1964) contended that, this being the case, it would be less deepened by glacial erosion than Western Passage. The fact that the two "valleys" are of similar depth "suggests that glacial erosion was not the principal agent" (Upson and Spencer, 1964, p. M33) in their formation.

The St. Croix bedrock valley (Western Passage) was the northernmost of several along the New England coast studied by Upson and Spencer (1964) who, because the depth of these valleys varied, concluded that they did not represent adjustment to lower sea-levels during glacial stages and , therefore, are not indicative of eustatic changes. The presence of a veneer of Wisconsin till indicates that the valleys were in existence before the Wisconsin ice reached the coast. Upson (1954, p. 295) stated that in the southern Passamaquoddy Bay area "the bedrock configuration fits a stream system that must have been formed before Wisconsin ice invasion and whose base level is now more than 400 feet below present sea level."

If the depth of the bedrock valleys is not related to a eustatic lowering of sea-level, the question arises as to why they are below sea -level at all. Upson and Spencer (1964, p. M36) suggested that either the coast "is still recovering from Wisconsin glacial loading or that the crust itself has been warped downward during Pleistocene and Recent time."

In the writer's opinion the depth of the St. Croix bedrock valley is best explained by a combination of factors: 1) the valley is pre -Wisconsin, cut to a sea-level below that which exists at present; 2) isostatic uplift is not complete; and 3) part of the depth is due to glacial deepening despite the similarity of depth between Western Passage and Head Harbour Passage.

The submarine topography of the Lower Saint John has received nothing like as much attention as that of Passamaquoddy Bay; nevertheless there are indications that the river and its tributaries once flowed to a lower level than at present. This was first recognized by Chalmers (1885, p. 14GG) who stated that there are indications of "a pre-glacial elevation of the region of 100 feet or more above that of the present day relative to sea level" and that "the depth of the Kennebeckasis and certain parts of the St. John Valley ... may be taken as indicating a still greater elevation than that given above."

Since Chalmers studied the area accurate hydrographic charts have been constructed, thus enabling a more definite statement on the situation. The greatest depths recorded in the various water bodies which constitute the lower Saint John system are listed in Table 3, all depths being given with lowest normal tide as datum.

Table 3 indicates depth only and it needs some elaboration. The hydrographic charts show the depth of water but they do not, of course, show the depth to bedrock which will be equal to, or greater than, the depth of water. The great depth of the Saint John River below the Reversing Falls can be explained by erosion caused by seaward fall of water at low-tide. The depth of 170 feet above the falls cannot really be explained in a similar way, although there is a landward rush of water in the river at high-tide. The greatest depth of water is not recorded in the "plunge pool" area, although the water is deep there (148 feet), but further upstream, north of Lee Cove. It was mentioned earlier that the Lee Cove - Manawagonish Cove route has been suggested as a former outlet for the Saint John River (Fig. 77). Another suggested former outlet for the river is from Drury Cove to Courtenay Bay. A deep-water channel (120 feet +) leads towards Drury Cove in the same way that a deep-water channel leads towards Lee Cove. In both cases it would be worth carrying out a seismic survey of the intervening land areas between the coves and the open sea, to see whether any evidence of buried channels can be found. A third possible former outlet is *via* South Bay, which is very shallow compared with Lee Cove and Drury Cove. However, it is now to one side of the main channel of the Saint John and its shallowness may be due to silting up since it was used as an outlet by the river (if it was used at all).

TABLE 3

MAXIMUM DEPTH OF WATER IN WATER BODIES IN THE LOWER SAINT JOHN VALLEY

Water Body	Hydrographic Chart No.	Depth (in feet)
Saint John River Mouth	4319	
Below Reversing Falls		170
Above Reversing Falls		170
Kennebecasis Bay	4344	204
Kennebecasis River	4344	194
Long Reach	4344	134
Belleisle Bay	4344	66
Washademoak Lake	4345	100+
Grand Lake	4362	100

In the north-east/south-west trending water bodies of the Lower Saint John Valley it is noticeable that depth does not increase steadily downstream. There is, instead, a series of basins and steps; for example, the greatest depth in Kennebecasis River (194 feet) is almost as great as that in Kennebecasis Bay (204 feet). These variations are difficult to explain without information about the nature of the bottom, but they may be due to unequal glacial erosion and deposition.

Melvin (1966) showed that in the Loch Lomond area, where the topographic trend is north-east/south-west, the distribution of glacial deposits is largely determined by the shape of the underlying bedrock surface. The same principle might apply in the whole Lower Saint John Valley and some of the shallow areas in the water bodies may represent areas of glacial deposition. It is noticeable that, except where the river becomes constricted towards the Reversing Falls, the depths in the north-west/south-east aligned part of the Saint John are much less than in Kennebecasis Bay and Long Reach. This suggests that, if ice erosion played a part in determining the depth of the water bodies of the Lower Saint John, deepening along a north-east/south-west axis was more effective than along a north-west/south-east axis. This situation may be compared with that of Head Harbour Passage and Western Passage.

It is well-established that the Saint John used to flow to a lower sea-level than it does at present, although ice almost certainly played some part in deepening the valleys. The problem of how much lower sea-level was and when lower sea-level occurred still exists. Even assuming that a substantial part of the depth of the water is due to glacial erosion of the river channel, the 100 feet suggested by Chalmers (1885) does not seem excessive; and, on the basis of the depth of 170 feet above the Reversing Falls, the writer suggests that 150 feet+ is a possibility. There is good evidence that the main features of the topography of the Lower Saint John Valley were in existence in pre-Pleistocene times, so it may be that the deep channels were eroded before glaciation and flooded to their present depth after it. Alternatively, they may have been eroded to their present depth during an inter-glacial, at a time when isostatic recovery was outstripping eustatic rise of sea-level, and then flooded in post-glacial time. It will be shown later that isostatic recovery of this area is probably not complete.

Some work has been done on tracing the bedrock surface in the Moncton area (Carr, 1961; Hobson and Carr, 1967) and it has been found that in the Petitcodiac and Memramcook river channels the elevation is well below sea-level. This indicates that before the deposition of surficial materials the base level of these streams, and consequently sea-level, was lower than at present. Exactly when this was so is not known, but Hobson and Carr (1967, p. 8) thought that "this ancestral drainage system probably extended well into the Bay of Fundy, and presumably was developed in post-Pennsylvanian and pre-Pleistocene time."

There is really no equivalent evidence of submergence on the Nova Scotia Fundy coast, unless Digby Gut (309 feet deep at its maximum) is considered. There is little doubt that Digby Gut is a drowned, fault -determined valley and that it existed in pre-glacial times as a gap across North Mountain. Because during glaciation it formed a convenient path for ice movement, it became deeper. When sea-level rose high enough relative to the land - possibly during interglacials - it formed a convenient channel through which water rushed into the Annapolis Lowland. This caused further deepening and, with an increase in tidal range, the process has continued to the present.

Evidence from the Continental Shelf

It can be shown conclusively that during and following the last glaciation sea-level was much lower in the Gulf of Maine and along the continental shelf off Newfoundland, Nova Scotia, and New England than it is now, and that the once-exposed landscape is now submerged. By implication, sea-level was then lower in the Bay of Fundy. The situation is complicated, however, because deglaciation came later in the Bay and the interplay of eustatic and isostatic changes is more difficult to decipher.

An interesting argument for believing that sea-level was lower along the Atlantic coast in post-glacial time than it is now was advanced by Fernald (1911). He found similarities between some elements of the flora and fauna of southern New England and those of Newfoundland and gave as a possible explanation the existence of a post-glacial land bridge extending along the offshore banks from New England to Newfoundland. The bridge would not have been continuous, being interrupted by Great South Channel, Northeast Channel and Laurentian Channel; but he did not believe that this would have been sufficient to stop the migration of the plants and animals concerned. He received some support for his explanation from Barrell (1915) but the latter pointed out that, using Fernald's explanation, the migration of flora would have taken place when the climate was cold,whereas the nature of the flora involved pointed to the need for warm conditions. Therefore, as an alternative to sea-level sinking, Barrell suggested a bulge peripheral to the ice sheet. However, some years later, Fernald (1933) rejected both possibilities and concluded that the migration pre-dated the glacial period.

Irrespective of whether the existence of a land bridge can be held to account for the similarity of flora in Newfoundland and New Jersey, it is still thought that an offshore ridge existed in post-glacial times; for example, Shepard, Trefethen, and Cohee (1934, p. 301) stated that "some evidence from the flora of Sable Island suggests that there may have been at one time a nearly continuous ridge of morainic material standing above water and extending from Sable Island to Long Island." However, the morainic origin is not completely accepted because Berger *et al. in* Garland (1966, p. 110) wrote: "It seems likely that the line of marginal banks from the Grand Banks of Newfoundland to Georges Bank is a depositional feature, related to pre-Pleistocene sedimentation rather than a series of morainic deposits."

During the 1960's attention was focussed on the finding below sea -level of animal and vegetable remains because, these can be radiocarbon dated, thus giving the time when sea-level was lower. Uchupi (1964b) reported the finding of a mastodon tooth on Georges Bank, but he was unable to decide whether it was transported there by ice or whether it was found near where the mastodon died. Medcof, Clarke, and Erskine (1965) found that oyster shells from Georges Bank and Northumberland Strait were 10,600 and 6,850 years old and, using Shepard's curve of eustatic change of sea-level (Fig. 57), calculated that when the specimens were alive sea -levels were between 132 feet and 50 feet below the present level. As the shells were taken from depths of 175 feet and 122 feet, by subtraction it was calculated that they lived at a depth of 43 feet and 72 feet. This is greater than the depth at which most oysters are found today and, on this basis, a subsidence of the coast independent of sea-level changes was postulated.

Emery, Wigley, and Rubin (1965) reported the finding of a sample of salt-marsh peat at a depth of 195 feet at the north-western margin of Georges Bank. As the peat contained rhizomes of *Spartina* (a salt-marsh grass which grows between mid- and high-tide), it was thought to indicate past sea-level better than shells. It also contained woody material which is considered to be more reliable for radiocarbon dating than shells. The date obtained for the sample was 11,000 ± 350 BP. Thus, about 11,000 years ago sea-level on Georges Bank was 195 feet lower than at present. Wigley (1966) thought that at this time Georges Bank was an ice-free island which he termed "St. Georges Island". The peat was thought to be *in situ* and, as it was found on the margin of Great South Channel (which was occupied by ice emptying from the Gulf of Maine during the late Pleistocene) it is of post-glacial origin. Eleven thousand years BP is, of course, a minimum date for the disappearence of ice; it may have been very much earlier. The much earlier dates for the disappearence of ice from the New Brunswick coast are not, therefore, in any way contradictory.

Whitmore, Jr. *et al.* (1967) claimed that the curve for sea-level rise obtained by study of dated shells and peat from the Atlantic Continental Shelf differs little from Shepard's (1963b) eustatic curve. Similarly, Emery *et al.* (1967) produced a curve based on radiocarbon dates from the Atlantic coast and shelf region showing that about 11,000 years ago sea-level was about 230 feet lower than present. It has been suggested that when this was so, man existed on what is now the continental shelf (Emery and Edwards, 1965-1966; Emery, 1966), although as far as the writer is aware no artefacts have been found to prove this. Since 11,000 years ago sea -level has risen rapidly until 7000 years ago and less rapidly since then (Emery *et al.*, 1967, Fig. 6).

In some parts of the world old shorelines below sea-level have been recognized. Not much evidence of this sort is available in or near the Bay of Fundy, but on the Scotian Shelf Yorath (1967) and King (1969) have identified a well-developed submarine terrace at a depth of 363-380 feet.

The conclusions drawn from the foregoing discussion of lower sea-levels is that, whether local (that is, the Atlantic continental shelf of eastern Canada and northern U.S.A.) or world-wide conditions are considered there is abundant evidence of lower sea-levels in post-glacial times; and that in both cases estimates of amount and timing vary.

Submerged Forests

The term "submerged forest" is in common usage but it can be misleading, in that it is sometimes used to describe a situation where no more than a dozen trees are involved. Here it is used to refer to any indication that trees which once existed above sea-level are now covered by sediments of marine origin.

The existence of "submerged forests" around the Fundy coast was known well over a century ago (Dawson, 1855a, 1855b). The known distribution of "submerged forests" around the Fundy coast is summarized in Table 4. Where two workers are listed as having reported on the same area, the two are included because there is some variation in what they reported. The table is probably not complete but includes all references that have been found in the literature (except in the cases where a person has referred to previous work) and three sites located in the field by the writer.

In the early days of the study of the submerged trees there was some disagreement about how the trees became submerged. Dawson (1855b, 1856) thought that the trees indicated a rise in sea-level but Hamilton (1868) argued that the drowned forests of Fort Lawrence are due to the erosive action of the tides, which undercut tree-covered land resulting in the slumping of slabs of land, with the trees still in place, to levels below sea-level. He received support for this "landslide" theory from Hind (1875).

TABLE 4

LOCATION OF SUBMERGED FORESTS IN NEW BRUNSWICK AND NOVA SCOTIA

Reported by	Date	Location	Depth	Remarks
Dawson	1855a	Fort Lawrence, between mouths of La Planche and Missaquash Rivers	"25 feet below level of highest tides" and "at least 30-35 feet below level of high tides"	Two separate levels. Pine (*Pinus strobus*) and beech (*Fagus ferruginea*) found
		Folly River, Chignecto Bay		
Gesner	1861	south-east coast, Grand Manan Island	18 feet	"The roots, stumps, and truhks of a great number of trees (the pine, hemlock, and cedar) still remain firmly attached to the sunken earth and at the very site where they flourished"
Baillairge	1874	Fort Lawrence, between mouth of La Planche and Missaquash Rivers	"10 feet at half tide"	Reference in *Baie Verte Canal*
Hind	1875	Fort Lawrence	"about 21 feet and 32 feet below the plane of the marshes"	The same levels identified by Dawson (1855a)
Matthew	1879	Navy Island, West Saint John	"between high and low water marks"	
Chalmers	1895	mouth of La Planche River	"30 feet below the level of the marsh or 80 feet below mean tide level"	These are depths to a "forest bed" and do not indicate the finding of trees *in situ*
		Aulac	"79 feet below mean tide level"	
		mouth of La Planche River	"10.8 feet and 0.3 feet below mean tide"	The depths refer to tree stumps *in situ* and were taken by Chalmers from *Acadian Geology*, supplement to Second Edition, page 13
		Edgett's Landing	"15.32 feet below mean tide level"	
Bailey	1898	head of St. Marys Bay	at least 5 feet below the marsh surface	Layer contained leaves, probably of birch or beech

TABLE 4 - Continued

Reported by	Date	Location	Depth	Remarks
Churchill	1923-1924	Boot Island		
Goldthwait	1924	Boot Island "the guzzle" Evangeline Beach	"33 feet below high tide"	
		Avonport, 2 miles south-east from Boot Island	"32 feet below high tide"	
		Hantsport		
		Yarmouth	"1 to 4 feet below the top of the marsh"	Located just outside the study area
Lyon and Goldthwait	1934	Grand Pré, "the guzzle"	17 to 31 feet below high water	
		Fort Lawrence	"near mid-tide level" and "about 5 feet lower"	
Crosby	1951 1962	mouth of Kennetcook River	"at least 20 feet below the high tide level"	Tree types are characteristic of dry upland rather than low or swampy ground
Cameron	1956	Avonport		Radiocarbon date 4,200 ± 200
Frankel and Crowl	1961	"near Amherst"		Radiocarbon date 5,300 ± 150
Carr	1961	Dieppe and Upper Coverdale (Petitcodiac Estuary)		
Langmaid	1968	Sackville, White Birch Marsh (Tantramar Marsh)	"beneath 1 foot of marine sediment"	Birch (*Betula*) tree with roots embedded in soil horizon. Radiocarbon date 640 ± 130 (GSC - 965)

TABLE 4 - Continued

Reported by	Date	Location	Depth	Remarks
MacNeill	1969	Tantramar Marsh, at Sackville control dam	22 feet below marsh level 25 feet below marsh level	Radiocarbon date 3,430 ± 110 Radiocarbon date 3,290 ± 100 MacNeill does not state whether stumps were found *in situ*. Samples dated included spruce, birch, spruce cones, and eel grass
Welsted	1969	Tantramar Marsh, northern part, 0.3 miles south of Large Lake	7 feet below marsh level	Radiocarbon date 935 ± 95 (I - 4312)
		Tantramar Marsh, in channel of Tantramar River, 3 miles north-east of Sackville	20 feet below marsh level	Radiocarbon date 1,875 ± 95 (I - 4313)
		Avon River, 0.7 miles north of junction of Kennetcook River	5 feet below high-water mark	Radiocarbon date 600 ± 95 (I - 4314). The same locality reported by Crosby (1951, 1962)
Taylor	1969	head of St. Mary Bay Church Point Meteghan River		Locations not exact - taken from Taylor (1969, Fig. 4)
Grant	1970[1]	Church Point	12 feet below "higher high water spring tides" (HHWST)	Radiocarbon dates 5060 ± 130 (GSC - 900)
		Grand Pré, opposite Boot Island	29 feet below HHWST	3820 ± 130 (GSC - 972)
		Grand Pré, opposite Boot Island	27 feet below HHWST	3840 ± 140 (GSC - 1054)
		Saints' Rest	8 feet below HHWST	8180 ± 150 (GSC - 757)
		Highland Village	17 feet below HHWST	2070 ± 130 (GSC - 957)

TABLE 4 - Continued

Reported by	Date	Location	Depth	Remarks
Grant	1970	Highland Village	12 feet below HHWST	Radiocarbon dates 1750 ± 130 (GSC - 1045)
		Lyon Head	6 feet below HHWST	1210 ± 140 (GSC - 973)
		Fort Beausejour	39 feet below HHWST	4040 ± 130 (GSC - 930)
		Fort Beausejour	31 feet below HHWST	3520 ± 140 (GSC - 975)
Langmaid	1970[2]	Fort Beausejour	13.72 feet below mean sea-level	White pine embedded in soil profile (podzol). Radiocarbon date 4120 ± 130 (GSC - 1089)

1, 2 In Lowdon and Blake, Jr., (1970)

It is now agreed that the idea of tree-covered land sliding or slumping to its present position below sea-level makes it impossible to explain the large number of trees that are found in a vertical position and seemingly undisturbed since the time that they were alive.

Therefore, the submerged trees are explained by a change in sea -level; but it is an over-simplification to say that they indicate a general rise of sea-level in the Bay of Fundy in recent times. There are three possible explanations.

First, there is the possibility of changed shore conditions, such as the breaching of a bay bar which would allow salt water into the bay; under storm conditions salt water might reach wooded areas killing off the trees and covering them by marine deposits. A detailed knowledge of the shore conditions of each particular locality is necessary before this explanation can be used with any certainty.

The second explanation is that there has been a eustatic rise of sea -level which has affected the whole Fundy coast. It is fairly generally accepted that, if any eustatic change is taking place at present, it is a rise rather than a fall (Donn and Shaw, 1963). However, in the Bay of Fundy this has to be balanced against the fact that isostatic uplift may not be complete and may be cancelling out any eustatic rise.

Third, the submergence of trees may represent not a general rise of sea-level but an increase in tidal range. As high-tide got higher, trees which were above sea-level were killed off and covered by silt deposited by the tide. In this connection, it is significant that the bulk of the "drowned forests" are located at the head of the Bay (particularly in

Cumberland Basin and Minas Basin) where the tidal range is greatest, and that there is a scarcity on the New Brunswick side where the range is least.

There is also the possibility of slow regional tectonism within the Maritime Provinces causing part of the subsidence. As Kupsch *in* Mayer-Oakes (1967) has stated, slow tectonic movements are likely to become more noticeable as post-glacial uplift slows down. While this possibility should not be ignored, detailed consideration of it is beyond the scope of this thesis.

Although the submerged trees have long been known to exist, only within relatively recent years (post 1950) has it been possible to date the time and rate of submergence with any certainty. Even so, it was clear to the early workers that the "rise of sea-level" took place some long time after deglaciation as some of the trees have roots in till or in peat lying on top of till. The first attempt at greater precision was by Lyon and Goldthwait (1934) who tried by tree-ring analysis to cross-date trees in drowned forests in New England and on the Fundy coast (at Fort Lawrence and Grand Pré). They were unable to make cross-identification and concluded "that submergence was so slow ... that a long time elapsed between the killing of one tree and the next" (Lyon and Goldthwait, 1934, p. 614).

With the advent of radiocarbon dating it became possible to draw curves of sea-level change based on the depth and age of submerged trees. Lyon and Harrison (1960) found that at Fort Lawrence and Grand Pré, Nova Scotia the rate of sea-level rise between 4500 and 3000 BP was three times as rapid as that at Odiorne Point, New Hampshire, where the rate of rise was close to the approximate average eustatic rise determined by Shepard and

Suess (1956). They suggested that the more rapid rise at the Nova Scotia sites was due either to crustal warping or to a progressive increase in tidal range, but they reserved making a decision until they had a greater number of dates. When they had carried out more precise surveys and had more radiocarbon dates, they published a more detailed assessment of changes of sea-level and crustal movements between 4500 and 3000 BP (Harrison and Lyon, 1963). The sequence of submergence and emergence that they suggested is complex and, in the writer's opinion, based on the fallacious assumption "that rising salt water killed the trees from which wood could be dated by C^{14} [and] that submergence of the forests was progressive at each site" (Lyon and Harrison, 1960, p. 295). Put another way, this means that if sea-level rises progressively the higher trees should be younger than the lower ones. If a tree in a higher position is older than one in a lower position (as some are in the sites investigated), it is necessary to resort to the explanation that there has been a regression followed by a transgression. This could be due to eustatic changes or to crustal warping; and because Shepard (1960) showed a steady eustatic rise during the period involved, Harrison and Lyon (1963) resorted to crustal warping as an explanation.

Gretener (1967) showed that the rare event in geology can be of great significance and the writer would rather resort to the effects of occasional severe storms as an explanation for the fact that in the drowned forests at Fort Lawrence and Grand Pré the highest tree is not always the youngest. There is good documentation of three severe storms during the past 250 years. These occurred on November 9, 1900 (Stead, 1903); on October 5, 1869 (the so-called "Saxby Storm"); and in November, 1759 (Hind, 1875). These storms caused the sea to overtop the dykes built by the Acadians at the head of the Bay of Fundy and, by flooding most of the Chignecto Isthmus, came close to converting Nova Scotia into an island. They resulted in sea water reaching wooded areas and killing trees.

Similar storms could have occurred between 4500 and 3000 BP and the following is a possible sequence of events. A severe storm killed trees up to a certain level. Following the storm, sea-level returned to its "normal" level and stayed about that level long enough for trees to regenerate in the once-flooded area. At a later date, another, but less severe, storm killed off trees up to a level lower than that reached by the first storm. Consequently, the dead trees at the higher level are older than those at the lower level. This pattern superimposed on a general rising trend of sea-level - which would result in the dead trees being covered by silt - could explain the situation at Fort Lawrence and Grand Pré and is preferred to the idea of complex crustal warping.

There is, however, no need to rely entirely on the altitude of individual trees. Study of other forms of buried organic matter and, of the general depth of silt below the marshes, allows a much more complete picture of submergence to be constructed.

The Marshes

Earlier it was observed that there are six major areas of tidal marsh around the Bay of Fundy. General descriptions of the marsh areas have been written by Ganong (1903), Johnson (1925) and Chapman (1960). The Tantramar-Aulac-Amherst marsh, located north of Cumberland Basin, is typical of the Fundy marshes. It occupies much of the southern part of Chignecto Isthmus and, because of the importance of the geomorphological history of the Isthmus already alluded to, it is given more attention in this account than the five other major marsh areas.

The Tantramar-Aulac-Amherst marsh is divided into 3 parts by Fort Cumberland Ridge and Fort Lawrence Ridge, which project about 100 feet above the general marsh level. West and north of Fort Cumberland Ridge the marshes are drained by the Tantramar and Aulac Rivers; the area between Fort Cumberland Ridge and Fort Lawrence Ridge is drained by the Missaguash; and south and east of Fort Lawrence Ridge drainage is by the La Planche River (Fig. 94).

It has been known for many years that the marshes of the Bay of Fundy are underlain by silt (marsh mud) which has been deposited by the tides (Dawson 1855a, pp. 23-31; Eaton, 1893) and accumulated as the sea-level rose or the tidal range increased. The rate of accumulation is amazingly swift: Chapman (1960, p. 41) quoted the deposition of 30 inches of silt in 122 days on the banks of the Avon River and the deposition of 1 inch by a single tide.

It was also known at an early date that the deposition of the silt beneath the marshes took place in post-glacial time (Trueman, 1899). If the silt were found resting directly on bedrock in all localities, there might be doubt about the post-glacial age of at least some of the silt. However, borings made for the purpose of locating the best route for a road to bypass the town of Amherst, Nova Scotia, reveal that the silt frequently rests on till.

Although it is generally agreed that the silt has accumulated as sea-level rose or tidal range increased, it has been suggested that, in the early stages of deposition, Cumberland Basin was a shallow fresh water lake (Monro, 1886; Ganong, 1903). Chalmers (1895, p. 128M) referred instead to a "quiet lagoon or recess" in this area, basing his reasoning on the fact that shallow water shells found in blue clay at Aulac are too well preserved to have been deposited in an area subjected to strong tidal currents such as exist today. Chalmers did not give the exact depth at which the shells were found, but recently shells (*Mya arenaria*) were found 17.3 feet below the marsh surface at a point 2 miles south of Aulac (Grant *in* Lowdon and Blake, Jr. , 1970) . They were found in growth position in salt marsh peat, suggesting flooding of the marsh surface by a temporary rise of sea-level, by an increase in tidal range, or by a change in coastal conditions. They have been dated at 1760 ± 140 (GSC - 1030).

The thickness of silt is great. Chalmers (1895, p. 41M) reported 80 feet of "marsh mud" in a boring at Aulac Station and more recently the Maritime Marsh Rehabilitation Administration (M.M.R.A.), in borings to determine the best location for the Tantramar control dam, found 96.9 feet of "blue clay" extending to a depth of -75.4 feet at the present site of the dam.[1]

As there were no records for locations further inland than those quoted above, the writer attempted to determine the thickness of silt at a number of localities using a hammer seismograph (Appendix 3). Traverses were made at 15 sites but at only 9 of them were results obtained which are of use in this study. It was found that the thickness of silt was in some cases greater than that for which the seismograph would give

[1]All elevations obtained by the M.M.R.A. and quoted in this thesis were obtained using Canadian Geodetic Datum. Therefore, unless otherwise stated, heights or depths refer to height above or depth below mean sea-level.

reliable readings. Therefore, most of the thicknesses given are minimum figures (Table 5). Where an exact figure is given, it records the depth from the surface to bedrock (mainly soft sandstone in this area) or to the top of the till. The location of the sites is shown in Figure 94.

TABLE 5

DEPTH OF SILT BELOW TANTRAMAR-AULAC-AMHERST MARSH
(Depths in feet)

Site No.	Depth	Comments
1	26	Sites located close to the coast
2	20	
3	40+	
4	3	Sites located well inland close to Fort Cumberland Ridge
5	5	
6	40+	Located in the area drained by Tantramar River
7	33+	
8	60+	
9	47+	

Information about the depth of silt beneath Amherst Marsh was obtained from the borings made in connection with the Amherst Bypass. In the records available to the writer the greatest thickness of silt (marsh mud) is 68 feet, recorded on the south bank of the La Planche River (unfortunately the height of the ground surface is not given); and the greatest depth at which silt is recorded is -38 feet at a point 2400 feet north of La Planche River.[1]

[1]The datum used was a bench mark whose height was determined using Canadian Geodetic Datum. Therefore - 38 feet is 38 feet below mean sea-level.

So far as the writer is aware the greatest recorded thickness of tidally deposited silt in the Tantramar-Aulac-Amherst marsh area is 96.9 feet, at the site of the Tantramar control dam; and the greatest depth at which it is recorded is 75.4 feet below mean sea-level, at the same location. However, there have been many borings in the general area in connection with various engineering works and it is impossible to gain access to them all; so evidence of thicknesses and depths in excess of those noted may exist.

The M.M.R.A. has provided information about the thickness of silt in some of the other marsh areas of the Fundy coast: near the mouth of the Nappan River, 0.3 miles north of Nappan Station, 37 feet of silt extends to a depth of -15 feet; at the mouth of Shepody River, at a point 600 feet east of the present control dam, 113.6 feet of marsh clay extends to a depth of -98 feet; and on the north bank of the Annapolis River, 0.7 miles north-east of Granville Ferry, 38 feet of "organic blue marsh mud" was found extending to a depth of -16 feet.

The total thickness of silt at any one locality does not indicate exactly the amount of rise of high-tide level because some of it, particularly the lower layers, was deposited below high-tide level. Also the complicating factor of compaction must be considered. The weight of sediment must have resulted in the compaction of the lower layers and a decrease in the total thickness of silt. That compaction has taken place is shown by one of the seismograph traverses (Fig. 95). The steady flattening of the curve, indicating increased speed of transmission of shock waves with depth, is due not to any drastic change of sub-surface material, but to increasing compaction with depth.

The problem of compaction is compounded by the fact that much of the so-called marsh is no longer natural marsh. Vast areas have been reclaimed for agricultural use, principally for growing hay. This is true especially in the Tantramar-Aulac-Amherst area.[1]

The old method of reclamation initiated by the Acadians was to build dykes along the river banks, to a height greater than highest high spring tides, and across the mouth of each small drainage channel leading to Cumberland basin to construct a simple but effective sluice (or aboideau). It consisted of a flap which was pushed open by water flowing seaward when the tide was low, but which was pushed shut by water attempting to move inland at high-tide. In this manner, and by the construction of artificial drainage ditches, the marshes were drained and the salt content gradually decreased. With the expulsion of the Acadians in 1755 many of the dykes fell into disrepair and some of the marshes returned to near natural conditions.

In the past 20 years, rather than try to preserve hundreds of miles of dykes, the M.M.R.A. has built dams across the major rivers, thus protecting the whole marsh area upstream. They are more elaborate than the old aboideaux, but the principle is the same; fresh water is allowed to flow seaward at low-tide, but salt water is excluded at high-tide. Such dams have been built across the Tantramar, Nappan, Shepody, and Annapolis rivers, completely transforming their nature; for example, although the Tantramar has a wide channel, it contains only a trickle of water and is never bank-full as it used to be at high-tide.

The drainage of the marshes, brought about by the old dyke and aboideau system and the modern dams and artificial drainage channels, has

[1]See Johnson (1925) for a description of the general nature of the reclaimed marshes.

resulted in the areas furthest away from the stream channels being at a lower altitude than the stream banks (Hind, 1875; Guilcher, 1958, p. 109).

A further complication arises as a result of the method used in the past to reclaim lakes and bogs north of Cumberland Basin (Harris, 1869; Godwin, 1893). The procedure was to cut a channel from the nearest tide water to the lake or bog. This would allow salt water to flow into the lake or bog at high-tide killing any vegetation and gradually burying it beneath marsh mud. When sufficient mud had accumulated, an aboideau was constructed to cut off the inflow of salt water but to allow seaward drainage. Within a few years the land was fit for agricultural use. It is not known when this procedure started but any peat or buried wood from the reclaimed marsh that has an age younger than the Acadian settlement of the area must be regarded with suspicion.

There are, then, really two types of "marsh": the reclaimed marshes protected from salt water by dykes and dams, and outside the dykes and dams the "natural" marshes which are periodically covered by salt water, the frequency depending on the height of the marshes. A clear distinction should be made between the two when discussing submergence.

Despite the complication involved it was decided to bore into the reclaimed parts of Tantramar, Aulac and Amherst marshes in the hope of locating peat layers at depth, as the finding of a peat layer between marine-deposited silt is regarded as a good indicator of marine transgressions and regressions (Jelgersma and Pannekoek, 1960). Salt-water peat forms at or close to high-tide level; thus, if a layer of salt-water peat is found beneath say 18 feet of silt, submergence of about 18 feet is indicated. Fresh-water peat is formed at an altitude greater than high-tide level; so, if a layer of it is found beneath 18 feet of silt, submergence of more than 18 feet is indicated. The existence of peat layers would indicate that the high-tide level has not risen continuously since the first silt was deposited - there must have been periods of still-stand (or regression), long enough for peat to form, and then renewed submergence. It was hoped, by boring through the silt, to locate peat bands and to obtain samples for radiocarbon dating and pollen analysis. Given the radiocarbon date for the peat and the depth at which it was found, it would be possible to work out the rate at which silt had accumulated; although, as the boring was being done in the reclaimed marsh area, the problem of compaction would have to be considered, both of the peat - unless the sample was taken from the base of the peat layer - and of the silt above.

The decision to concentrate on the reclaimed marsh rather than on the natural marsh was based on practical rather than academic reasoning. The reclaimed marshes in dry weather are easy to travel, although they can be exceedingly difficult in wet weather. The natural marshes are always difficult to cross and a small boat is needed (a miniature Hovercraft would be ideal); also, of course, the worker is entirely dependent on the state of the tide.

The instrument first used for the borings was a Jarret auger, with which samples were obtained from depths as much as 22 feet below the marsh surface. Although peat layers were penetrated, none of them was very thick and it was not possible to obtain a large enough sample for radiocarbon dating. Later an instrument known as an hydraulic porta-sampler was used. The hydraulic mechanism facilitated penetration to 39.5 feet below marsh level at one site (18.9 feet below mean sea-level), but the instrument is not ideal. Again, peat layers were encountered in the borings but it was not possible to collect enough peat for radiocarbon dating.

In all, 35 boreholes were sunk and in 14 cases peat or other organic material was found (Fig. 96). Ten samples collected from the boreholes, together with two peat samples dug from the side of the Tantramar River channel, were subjected to pollen analysis (Appendix 4). The analyses show conclusively that the levels at which the specimens were collected were once above high-tide level and that, since their deposition, high-tide level has risen at least to the level of the marsh surface below which the specimens were collected. There is, of course, the possibility that the high-tide level was "artificially raised" by settlers during the process of reclaiming lakes and bogs. It is unfortunate that no radiocarbon date is available for any of the specimens, but in the case of sample 11 (Appendix 4) some indication of age can be obtained. This sample was collected from the side of the Tantramar River channel about 7 feet below the general marsh surface (Fig. 97). A wood stump from the bottom of the channel (Fig. 98), approximately 1000 feet downstream, has been dated at 1875 ± 95 BP (I - 4313). The channel bottom here is approximately 20 feet below marsh level which is virtually horizontal in this area. The wood stump is, therefore, at a lower level than the peat band and, on this basis, it is suggested that it is the older of the two. The peat band is, therefore, less than about 1875 years old.

Crabtree (1968) explained the usefulness of pollen analysis in establishing stratigraphical correlations and suggested (personal communication) the possible correlation of two of the pollen samples with stages of the pollen sequence at Gillis Lake, Nova Scotia (Appendix 4). However, because of the difficulty experienced by the writer in obtaining uncontaminated samples, this line of reasoning is not pursued here although it is certainly one which deserves more detailed research and attention.

The greatest depth at which the writer found peat in the Tantramar Marsh area was 38.5 feet below marsh level at a depth of 17.8 feet below mean sea-level at site 1 (Fig. 94). In the literature the greatest depth recorded appears to be 80 feet below marsh surface at Aulac recorded by Chalmers (1895)[1] who attributed this great depth to subsidence along a nearby fault. However, the writer doubts whether there is any need to resort to faulting as an explanation.

Some further evidence of organic deposits within the silt is forthcoming from borings for engineering projects. Those done for the Amherst Bypass extend in a roughly north-south line across La Planche River west of Amherst. The borehole records make frequent reference to "organic silt and clay"; peat layers are sometimes referred to, although no precise depth is given for them. It is noticeable that in many of the borings the sequence is marsh mud (organic silt and clay), sand and gravel, till, bedrock. This suggests that, after ice disappeared from the locality, outwash deposits were laid down or that marine deposition took place in an environment very different from that which exists in Cumberland Basin today. If the latter was the case, deposition of marsh mud must have started long after deglaciation. Without detailed study of the sands and gravels it is not possible to say which of the two possibilities is the more likely.

Borings made in preparation for the construction of the Shepody Dam include 5 which record "marsh clay with considerable organic material", to

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[1]No figure is given for the relationship to sea-level.

depths of 82, 98, 98, 104, and 109 feet below the marsh surface;and, more specifically, one borehole records "heavy organic material" at 84.4 feet below mean sea-level, 100 feet below the marsh surface. Unfortunately, engineers are not usually interested in unravelling the past and no samples were collected for radiocarbon dating.

Other evidence of submergence

Several other lines of study could be pursued in order to determine whether or not submergence has taken place and is taking place, and some pieces of evidence are available which would deserve detailed study. In the main, they are concerned with the recent past rather than with several thousand years ago.

Bailey (1898) reported that at the head of St. Marys Bay oyster shells were found beneath five feet of soft black mud. He reasoned that, as oysters had not lived in St. Marys Bay for "forty or fifty years", there had been a change in climate and a depression of the land. The situation indicated that submergence had taken place, the accumulation of the mud occurring in a <u>minimum</u> or 40 or 50 years. If the shells could be located now, a radiocarbon date for them would add to the knowledge of the submergence of this part of the coast. However, a radiocarbon date of 760 ± 130 (GSC - 997) for rhizomes of *Scirpus sp.* taken from under 6 feet of high-tide salt-marsh gives some idea of the rate of submergence of the area.

Shell-heaps left behind at Indian camps are being eroded by the sea and it has been claimed that this is evidence that submergence is continuing (Bailey, 1897; Ells, 1907). Certainly there are several examples in the area surrounding Passamaquoddy Bay and study of them might reveal information about submergence. However, each site would have to be mapped in detail and local shore conditions would have to be taken into account in order to ensure that the "submergence" is not due to normal coastal erosion.

Of a similar nature is "an old Indian encampment" reported by Cameron (1956b, p. 7) in the Gaspereau Valley. The site is now below high -tide level and if a radiocarbon date were obtained for charcoal found at the site it would give the date of occupation and an idea of the rate of sea-level rise. Erskine (1960, p. 340) reported a beach site at Bear River which, he presumed, originally had a seaward projection now below high -tide level and, therefore, indicates a rise of sea-level.

Historical evidence can be of some value. Ganong (1901) claimed that Dochet Island in the St. Croix River was being submerged. His reasoning was that a ledge which used to be covered by trees is now (1901) bare. Later, he published a much more detailed description of the island (Ganong, 1902b) and, on the basis of a comparison of several maps, concluded that there was evidence of submergence at a rate of between one and two feet per century.[1] North of Cumberland Basin a small stream, La Coupe River, flows to Aulac River which drains south to the Basin (Fig. 94). Situated on La Coupe River is a system of embankments referred to in the area as an old French dry dock (Webster, 1933). La Coupe River, now no more than a trickle, would not support a small row boat. If ships large enough to sail in the Bay of Fundy did reach the "dry dock", La Coupe River has been reduced drastically in size, probably as a result of deposition of tide -transported silt. This could indicate a rise in high-tide level, or that

[1] Several maps of this small and seemingly insignificant island exist, because Champlain spent his first winter in North America on it - one map dates back to 1604. Also, the island was involved in a boundary dispute between Canada and the United States, resulting in more maps being drawn.

the drainage to the dock was not maintained. At Grand Pré artifacts discarded on the surface of salt marshes have been buried by tidal mud and near Fort Beausejour a corduroy road originally laid across the marsh surface is now overlain by 3 feet of high-tide salt-marsh mud (Grant 1970).

Finally, some information about whether submergence is taking place should be available from tide gauge records. However, only one tide gauge in the area has records for an extended period of time - Saint John, New Brunswick. In addition, tide gauge records are notoriously difficult to decipher because of changing coastal conditions. However, the tide records for Saint John show that since 1934 mean-tide level has been increasing.[1]

Amount, Timing, and Cause of Recent Submergence

The occurrence of wood stumps below high-tide level and of peat and other organic remains below tidally deposited silt clearly indicates that submergence has taken place. The points in question here are: 1) how much submergence has taken place and when did it occur? and 2) what was the cause of the submergence - general increase in sea-level or an increase in tidal range?

One of the earliest estimates of the amount of submergence was by Upham (1895) who suggested 80 feet at the head of the Bay of Fundy. Possibly this estimate was based on the report by Chalmers (1895) of peat 80 feet below the marsh surface at Aulac. The peat bed referred to was thick (20 feet) and it must have been formed close to or above high-tide level. The marsh surface at the borehole represents high-tide level at least, unless

[1]Details can be found on page 331 of *Water Levels, 1966: Volume 2 - Tidal*, Canadian Hydrographic Service, Ottawa, 1967.

it is located within the reclaimed marsh area, in which case compaction of the peat and silt may have resulted in its being lower than high-tide. The stratigraphic evidence at Aulac is taken by the writer to indicate at least 80 feet of submergence at the head of the Bay of Fundy. The maximum figure verified by the writer is at least 38.5 feet at site 11 on Tantramar Marsh (Figs. 94 and 96). However, it is considered that this represents a maximum based on the limitations of the methods used rather than a true maximum for submergence.

The record of borings at Shepody Dam, where "heavy organic material" is found 100 feet below the marsh surface, may indicate an increase of high-tide level as great as 100 feet in this area. Unfortunately, the nature of the organic material is not specified and it could represent material floated into place and deposited well below high-tide level. The record here is regarded as good circumstantial evidence rather than proof of the amount of submergence.

At the time of writing, the lowest tree stump found *in situ* is at Fort Beausejour at a depth of 39 feet below "Higher High Water at 'large tides'" (Grant 1970)-a datum which can be taken as representing the highest part of the salt-marsh surface in its natural state. This stump has been dated at 4040 ± 130 (GSC - 930) and clearly indicates that the highest level reached by the sea at this point has risen by 39 feet in about 4000 years.

The question of when submergence occurred is complex. It is thought that the Bay of Fundy, with the exception of Passamaquoddy Bay, was ice free by 13,000 BP while Chignecto Isthmus was ice free by 12,500 BP. It is known, however, that at 13,000 BP sea-level was at least 130 feet higher than at present in the southern part of the Bay of Fundy (the Pennfield Plain area) and, as the Bay is roughly parallel to the "isobases" for the area, it was probably 130 feet higher at the northern end of the Bay as well. This means that Chignecto Isthmus would almost certainly have been flooded at some time between 13,000 and 12,500 BP. However, archaeological evidence from north of Minas Basin indicates that the Isthmus was above sea-level by 10,585 ± 47 years BP (Borns, 1966). This is the average date for 13 artifacts found at Debert Paleo-Indian site (Stuckenrath, 1966). It is suggested that the people must have come to Nova Scotia from mainland North America. It is unlikely that they could have travelled along Georges Bank and Browns Bank, even if the banks were above sea-level, because Great South Channel and Northeast Channel would have presented too great an obstacle to people without boats (Byers, 1966). Therefore, their most likely route was *via* Chignecto Isthmus. The emergence which made this possible may have occurred at the same time as the development of a local ice cap on South Mountain which is tentatively correlated by some workers with a widespread readvance of ice (the Valders) in other parts of North America. The distribution of ice at that time must have been limited to certain localities or it would have cut off the migration path of the Debert people, and clearly it did not cover Debert.

During the time period when Chignecto Isthmus was ice free but beneath sea-level, marine deposition would have occurred and, in the early stages when ice was close, the sediments would probably have been coarse textured. This may account for the sand and gravel which occurs so frequently above till and below marsh mud in the borings in the La Planche Valley.

When the Isthmus rose above sea-level, some of the till and marine deposits were probably eroded by streams flowing to the Bay of Fundy. This resulted in irregular distribution of these deposits so that now they are not always found beneath the tidally deposited marsh mud, which in some areas rests directly on bedrock.

Using Shepard's (1963) curve of eustatic changes (Fig. 57) world sea-level 10,600 years ago was about 110 feet lower than it is at present; but this does not mean that sea-level was necessarily 110 feet lower relative to Chignecto Isthmus than it is today, because isostatic recovery of the latter would have been taking place. However, if Shepard's (1963b) curve is correct (and it is regarded by the writer as the most reliable yet published) sea-level has risen more or less continuously in the past 10,600 years and it has not been higher at any stage than it is now , although it may have reached its present level 2000 years ago (Shepard and Curray, 1967) (Fig. 58). Whether Chignecto Isthmus has been completely below sea-level at any time during the last 10,600 years depends on the balance between isostatic uplift of the land and eustatic rise of sea-level. McRoberts (1968) claimed that the Isthmus was flooded between 11,800 and 6400 BP, but this conflicts with the Debert evidence.

As sea-level rose it flooded the Gulf of Maine and the deeper parts of the Bay of Fundy but, in the early stages of submergence, the geometry of the Bay was different from that of today and the tidal range was consequently less (Rao, 1968). Thus, the early stages of submergence were the result of a straightforward rise of sea-level. However, as the Bay began to approach its present shape, the tidal amplitude began to increase. Grant (1970) suggested that the process began about 6000 years ago and he explained much of the submergence since then as a result of increased tidal amplitude. Basing his opinion on meticulous study of the relative rates of submergence of the Atlantic and Fundy coasts of Nova Scotia, he found that over the last 4000 years high-tide level in the Bay of Fundy has risen at a rate of about one foot per century, whereas along the Atlantic coast the rise over the past 2000 years has been about 0.5 feet per century. The rates are greater than the eustatic rise of sea -level by factors of 5 and 2.5 respectively.[1] He explained the exaggerated submergence along the Atlantic coast as the result of eustatic rise of sea-level producing hydro-isostatic subsidence of the continental shelf. In the Bay of Fundy the eustatic rise of sea-level gradually produced the present geometry of the Bay, with the result that tidal amplitude increased. By 8000 years ago sea-level had overtopped Georges Bank and Browns Bank but not until 6000 years ago was water at the entrance to the Gulf of Maine deep enough to allow tidal amplification in the Bay of Fundy. Moreover, to begin with, tidal amplification was limited and it was not until 4000 years ago that a rapid increase in tidal range began. Grant appears to have considered only eustatic changes and to have ignored isostatic changes in arriving at these conclusions. This may be justified in the case of Georges Bank and Browns Bank which were near the limit of the last ice sheet; but the Gulf of Maine and the Bay of Fundy were covered by ice during the Wisconsin, and isostatic effects should be considered.

Andrews (1970, Figs. 7-8 and 7-9) has constructed isobase maps for northern and eastern North America which attempt to estimate the amount of emergence over the last 8000 years and 6000 years. The map for the last

[1]Based on eustatic rise of sea-level determined by Shepard and Curray (1967) and modified by Scholl, Craighead, and Stuiver (1969) for the last 3000 years.

8000 years shows the zero isobase crossing the Bay of Fundy in a north -south direction, the head of the Bay lying outside the zero isobase; and the map for the last 6000 years shows most of the Bay, with the exception of the New Brunswick side south of Saint John, outside the zero isobase. As Shepard's (1963b) eustatic curve indicates that the rise of sea-level between 8000 BP and 6000 BP was at a rate of about 0.5 meters per hundred years and that sea-level has continued to rise from 6000 BP to the present, it is to be expected that submergence of the Fundy coast would take place beginning at about 8000 years BP. Andrews' maps are only "models" but the fact that they correspond well with field evidence suggests that they are essentially correct for southeastern Canada.

As high-tide level increased, plants which were once above sea-level were inundated and submerged. Silt deposited by the tide would have continued to accumulate on the marshes as long as there was no human interference. Hydro-isostatic depression was probably not a factor in the rapid submergence of the head of the Bay of Fundy, because the depth, total volume, and weight of water involved was limited. The writer has considered the possibility that the great accumulation of silt has contributed to gradual depression of the crust in the manner suggested for the Mississippi Delta (King, 1965) but has eliminated it as a possibility because most of the sediment deposited has quite clearly been eroded from the outcrops of soft Triassic sandstone along the coast of the head of the Bay. Thus, in effect, the total crustal load has not changed; it has merely been converted from one form to another. This is quite different from the Mississippi situation in which the sediment of the delta has been derived from the whole drainage basin and concentrated in one locality.

In Grant's opinion (personal communication) the geometry of the Bay has now reached, and may have passed, the optimum for tidal amplification with tidal range now declining. The charts produced by Gutenberg (1941), Moore (1948) and Valentin (1954), all of which are based on tide gauge records, show greater submergence of the Atlantic coast of Nova Scotia than of the Fundy coast in recent years. This suggests that Grant may be correct and that hydro-isostatic depression of the Atlantic coast is continuing, while submergence of the Fundy Coast has slowed down or stopped because the tidal amplitude has reached or passed its maximum. There is also the possibility of slow, long-term, regional tectonic activity isolated from isostatic and eustatic changes resulting in changes of relative sea level.

As was indicated earlier the rise of high-tide level within the Bay of Fundy was not a continuous process. The existence of peat bands between tidally deposited silt beneath Tantramar Marsh indicates that there must have been periods when high-tide level fell and then rose again. However, the fact that the peat bands located were all thin suggests that the interruptions were not of long duration.

Local conditions - impossible to deduce from boreholes alone - would have played a part in the preservation of trees and peat, thus complicating the problem of interpreting the rate and cause of submergence; for example, MacNeil (1969b) reported finding wood at the Tantramar Dam at depths of 22 feet and 25 feet below marsh level. He gave the impression that the lower of the two samples was the younger (3290 ± 100 as opposed to 3430 ± 110). This situation could have resulted from the shifting of the channel of the Tantramar River as sea-level rose.

Summary and Attempted Chronology of Quaternary Events

At one stage during the Pleistocene ice covered the whole of New Brunswick and Nova Scotia, but little is known of pre-Wisconsin events. Grant (1968) reported pre-Classical Wisconsin shell-bearing tills along the Nova Scotia Fundy coast north of Yarmouth. The age of these tills is supported by two radiocarbon dates of >38,000 (GSC - 695) and >39,000 (GSC - 887). There is, of course, no way of telling how much older than 38,000 and 39,000 years they are, or how long before the Classical Wisconsin they were deposited. Other tills are found in a higher stratigraphic position, so clearly there have been at least two phases of glaciation; but without detailed study of the problem it is pointless to speculate on their magnitude and extent.

Other workers have found evidence of more than one glaciation, but usually this evidence is very local in character and can be explained by minor fluctuations of the ice margin; for example, Wickenden (1941) found three types of till in an exposure at Joggins which "may indicate three distinct ages of glaciation, or advances of ice from different directions" (Wickenden, 1941, p. 148). Hughes (1957) reported two tills at Milford Station, but they "are believed to indicate a minor fluctuation of the ice margin rather than a non-glacial interval" (Hughes, 1957, p. 7). At Weymouth, MacNeil (1961, p. 5) found "three till sheets tentatively ascribed to the Tazewellian, the Caryan, and the Wisconsin (Valders or Cochrane) substages". On the basis of the evidence presented this chronology must be regarded as questionable.

Nevertheless, it is known that - at the glacial maximum - ice covered the whole of the two provinces, reaching as far as the Scotian Shelf. It has been traditionally assumed that the general movement was from north-west to south-east, although this concept has recently been challenged (Prest and Grant, 1969). Longitudinal movement of ice down the Bay of Fundy was proposed long ago and there is evidence that such movement did take place.

The disappearance of the ice from the region was not brought about by parallel retreat to the north-west but by the splitting of ice into separate segments caused by eustatic rise of sea-level flooding the Bay of Fundy and its offshoots. Once "New Brunswick ice" and "Nova Scotia ice" had been separated in this way, the ice front may have retreated north-west across New Brunswick leaving behind moraine ridges as it did so (Gadd, 1970a). At a later date ice re-formed on South Mountain,and possibly the Cobequid Mountains, although Prest and Grant (1969) regarded these ice masses as remnants of the former ice sheet that covered Nova Scotia rather than as newly formed ice caps. Whatever the explanation, the result in the case of the South Mountain was the movement of ice and morainic debris north-west towards North Mountain and the Fundy coast.

With the exception of Passamaquoddy Bay, which was occupied by an ice lobe until about 12,300 BP, the Bay of Fundy was free of ice by 13,000 BP. At that time sea-level was higher relative to the coast than it is now and there was, on the New Brunswick side at least, a transgressive phase of deposition as the ice margin retreated. Some indicators of this period of higher sea-level have been left behind, the most reliable being marine shells. Unfortunately, they give only a minimum figure for the amount of emergence that has taken place. Using this principle as a basis for calculation, it is known that 13,000 years ago sea-level in southern New Brunswick was at least 130 feet higher relative to the land than it is at present. It is probable that some time before 13,000 BP sea-level was as much as 250 feet higher than at present.

Although scarcity of shells on the Nova Scotia coast makes the identification of higher sea-levels less certain, there is good evidence that relative sea-level was once 137 feet higher than at present. On the whole, the figures obtained for Nova Scotia are lower than those for New Brunswick, suggesting up-tilt to the north-west; but, in the absence of any date for Nova Scotia, it is impossible to construct an isobase map.

Evidence of coastal submergence is abundant. Coastal configuration and river-bottom profiles show that submergence has occurred, but it is difficult to determine both time and amount of submergence using these bases, although an estimate of 150 feet is given for the Saint John Valley. Information obtained from below sea-level in the Gulf of Maine and on the Atlantic Banks indicates that at one stage much of the Bay of Fundy must have been above sea-level. Along the coast, submerged tree stumps and other submerged organic materials have been radiocarbon dated and, by accurate levelling, the amount and rate of submergence have been determined. There may have been 80 feet of submergence at the north end of Cumberland Basin and there has certainly been 39 feet in the last 4000 years. Other indicators of submergence could yield useful evidence of recent changes, but detailed surveying and careful appraisal of each site would be necessary in order to obtain reliable results.

Within the Bay there has been an interplay of isostatic changes of the earth's crust and eustatic changes of sea-level. Following deglaciation Chignecto Isthmus must have been flooded. However, archaeological evidence suggests that it had emerged by about 10,600 years ago, so there must have been a period of regression in the Bay of Fundy between 13,000 and 10,600 years ago. This accords with the conclusion by Andrews (1970), based on observations in Arctic Canada, that post-glacial uplift is rapid in the early stages. Whether the Isthmus has been flooded during the past 10,600 years is debatable and depends on the balance between isostatic changes in the area and world-wide eustatic changes. However, it seems that since about 6000 years ago the tidal range within the Bay of Fundy has been increasing, especially during the last 4000 years, resulting in the drowning of trees and peat layers. As it is unlikely that the tidal range would increase if the Isthmus were flooded, it must have been above sea-level during the past 6000 years. There are some indications that the tidal range has reached its maximum and that it will now decline, presumably resulting in the emergence of some features that have been drowned. Moreover, Andrews (1970, Fig. 9-5) showed that except for Minas Basin and Cobequid Bay there is over 40 meters (132 feet) of residual uplift (Urr) in the Bay of Fundy, and the same author has calculated (1970, Table V-1) that the rate of present uplift at Saint John is 0.330 meters/100 years (approximately 1.01 feet/100 years).

CHAPTER 6

SUMMARY AND CONCLUSIONS

Apart from some local exceptions the Bay of Fundy becomes deeper and wider to the south-west, towards the Gulf of Maine. These two water bodies form part of the North Atlantic continental shelf and are separated from the continental slope by Georges Bank and Browns Bank. The geometrical proportions of the Bay of Fundy are such that it has the greatest known tidal range in the world.

Rocks exposed along the Fundy coast range in age from Precambrian to Recent, although there is a gap between the end of the Triassic and the Pleistocene. Variations in lithology, and consequently variations in resistance to erosion, are responsible for some of the major features of the outline of the coast. For example, Minas Passage, only 3 miles wide between the "hook" of North Mountain and Cape Sharp, is developed between basalt outcrops; whereas Minas Basin to the east, developed on the non-resistant rocks of the Annapolis Formation, is a major southward embayment of the coast. Rock type, structure, exposure, and geomorphological history (including changes of sea-level) have combined to produce a great variety of coastal types, one of the most distinctive features of the Bay being the extensive development of tidal marshes and mud-flats.

The Fundy trough was in existence by the Triassic although it was not necessarily below sea-level. Sediments were deposited in the trough but at times sedimentation was interrupted by volcanic activity. Later, major drainage lines were developed along what are now the Bay of Fundy and the Annapolis-Cornwallis Valley. These two lowlands were later connected to each other by a lowland developed along a fault (one of several) which cuts across the North Mountain Basalt. The timing of these events is not certain, but they must have occurred between the end of the Triassic and the beginning of the Pleistocene.

Almost as soon as Powell elucidated the concept of the peneplain and Davis gave it a name, one was identified in Nova Scotia and New Brunswick. It was supposed to have been formed during, or at least by the end of, the Cretaceous Period. Practical tests show that, even if the peneplain concept is regarded as valid, there is no evidence that such a landform developed in New Brunswick and Nova Scotia. A discussion based on the facts now available leads to the same conclusion. An erosion surface, consisting of a series of steps rather than a gentle slope, truncates the rocks of the Atlantic Uplands of Nova Scotia; but it is believed to be a surface of pre-Triassic age. There is no evidence that it is genetically connected with other surfaces in Nova Scotia or New Brunswick.

The best known epochs are the Pleistocene and the Recent. The sequence of glaciation and deglaciation of the area is particularly important in this study because it complicates the deciphering of evidence of sea-level changes. It is known that at the glacial maximum New Brunswick and Nova Scotia and the Bay of Fundy between them were covered by ice. There is evidence of an earlier pre-Classical Wisconsin glaciation, but little other evidence of multiple glaciations of the whole area. It is thought that during deglaciation the ice sheet split along the Bay of Fundy and its offshoots, and that either a local ice cap formed on South

Mountain or a portion of the ice sheet remained. A similar situation may have occurred on the Cobequid Mountains but the evidence is less conclusive.

At the time of deglaciation, or very soon thereafter, relative sea -level along the southern New Brunswick coast was at least 130 feet higher than at present. This figure is regarded as certain because shells are found at this height. In addition, there is evidence that the sea in southern New Brunswick was probably at some stage 250 feet above its present level. On the Nova Scotia coast the maximum figure recorded is 137 feet, which suggests up-tilting of the crust in a north-westerly direction, although the tilting is not certain, since the 250-foot level and the 137-foot level are not necessarily synchronous. The figures quoted are for the marine limit, no correction having been made for eustatic change.

Submergence of the Fundy coast is indicated in places by its outline, by the depth of river channels, by information from the continental shelf, by the existence of buried organic materials, and possibly by some historical evidence. The amount of submergence indicated by the first two lines of evidence is difficult to measure, but a figure of at least 150 feet is suggested, although exactly when sea-level was 150 feet lower than at present is not known. In the case of buried organic materials dating is easier, a figure of -39 feet for 4040 years ago having been recorded.

During the Recent Epoch there has been, in the Bay of Fundy, an interplay between isostatic forces which have tended to cause a decrease in relative sea-level and eustatic changes which have tended to cause an increase in relative sea-level. With the information available at the time of writing it seems that the Bay was ice free by about 13,000 BP and that Chignecto Isthmus was uncovered by 12,500 BP. At this time relative sea-level was high enough to submerge the Isthmus, resulting in tidal conditions very different from those of today. However, archaeological evidence suggests that by 10,600 BP the Isthmus was above sea-level, indicating that in this area isostatic rebound was outstripping eustatic rise. What happened immediately after 10,600 BP is not clear, but there is evidence that over the last 6000 years the tidal range in the Bay has been increasing, suggesting that the Isthmus has been above sea-level for most of that time. The increase in tidal range is thought to have been responsible for the submergence of many of the forests and peat layers around the Bay.

The present trend is not known for sure, but there is some evidence that isostatic uplift is not complete; thus, further emergence might be expected. Furthermore, the tidal range in the Bay of Fundy may have reached or passed its maximum, in which case the high-tide level will decrease, giving the impression of emergence. However, the only long-term tide gauge record available, that for Saint John, indicates an increase in mean-tide level since 1934.

The writer has tried to describe the morphology of the Fundy coast and of adjacent areas and has attempted to explain the evolution of the Bay from the Triassic to the present, the emphasis being on Quaternary events. In the sections on sea-level changes the difficulties of interpretation have been stressed as has the need for a realistic statement of the degree of accuracy involved in any assessment of changing sea-levels.

As was stressed in Chapter 1 this is a general account, but as work has progressed the need for detailed study of some key areas and topics has become apparent. Some such studies have already taken place and the results are incorporated into this thesis. However, there are several others which would help towards a fuller understanding of the evolution of the Bay of Fundy:

(1) Study of the relationship between contemporary coastal morphology and the various tidal levels within the Bay is required so that increased accuracy in measuring sea-level changes may be achieved.

(2) As the glacial history of the area is intricately interwoven with sea -level changes, a comprehensive study of the glaciation and deglaciation is required. Most in need of attention are the Saint John area, the area between Saint John and Moncton, and that north of Shepody Bay and Cumberland Basin. In the Saint John area this would involve a thorough study of the various outlets of the Saint John River.

(3) Although the writer tried to determine the geomorphological history of the Chignecto Isthmus, many uncertainties about its evolution still exist. A method needs to be devised whereby uncontaminated peat samples large enough for radiocarbon dating can be taken from beneath the marshes.

(4) In some isolated areas of the coast terraces which may be of marine origin show up on air-photographs. These need to be looked at and surveyed in the field.

(5) In the case of some features now above sea-level but thought by the writer to be of marine origin, the heighting is imprecise. Accurate heighting of these features would permit more definite statements about the amount of emergence indicated.

(6) Pennfield Plain may have been washed by the sea when relative sea-level was higher than it is now. Fleming (1965) has claimed that, if sea-level is stationary, there is a maximum terrace width depending on the seaward slope of the land. Using this as a basis, it would be worth levelling across the Plain from north to south in order to get some idea of whether sea-level could have stood at the height of the Plain for any length of time.

(7) Isostatic recovery of the New Brunswick Fundy coast may not be complete. Therefore, trying to calculate the amount of residual depression using recent studies in the Canadian Arctic as a model (Andrews, 1968a, 1970) might be of some value.

Even if all the studies suggested are completed, there will be no one answer to some of the questions raised in this thesis. In the past the tendency has been to oversimplify things and this will probably continue however much information is collected, for as Bates (1964, p. 157) said:

> Man likes to simplify things, to find single causes to find an order in nature that corresponds with an orderly arrangement of ideas in his own mind.... But nature is...frightfully complex, perhaps too complex ever to be "understood" through the processes of our limited brains - and our fondness for single causes has probably got us in trouble more often that it has helped us.

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